Milton Updegraff
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February 1919.

INTRODUCTORY METEOROLOGY.

INTRODUCTORY METEOROLOGY

PREPARED AND ISSUED
UNDER THE AUSPICES OF THE
DIVISION OF GEOLOGY AND GEOGRAPHY
NATIONAL RESEARCH COUNCIL

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PREFACE.

Meteorology is included in the course of study outlined by the Committee on Education and Special Training of the War Department for Students' Army Training Units. The plan involves an intensive study of the elements of the subject in order to familiarize prospective Army officers with its chief conclusions and methods. To meet this requirement Introductory Meteorology has been prepared and issued under the auspices of the National Research Council with a view to its use in connection with a textbook on Military Geology. It is recognized that the book is inadequate for the needs of the professional meteorologists in civil or military service; training in physics and mathematics is essential in the interpretation of the causes of meteorological phenomena. On the other hand the book is not intended to be elementary in the sense that it needs no study on the part of students or elaboration and explanation by the instructor. Instructors will profit by a study of the books listed in the Bibliography and by consultation with the officials of the local Weather Bureaus.

The manuscript has been prepared by the staff of the United States Weather Bureau as follows: Chapters I, V, VI, VII and VIII by Dr. W. J. Humphreys, Professor of Meteorological Physics. Chapter II by S. P. Fergusson, in charge of Division of Tests and Repairs. Chapters III and IV by W. R. Gregg in charge of Aerological Investigations. Chapter IX by J. Warren Smith, Chief, Division of Agricultural Meteorology. Chapter X by A. J. Henry, Meteorologist and Forecaster. Chapter XI and Bibliography by Professor C. F. Talman, Librarian. Valuable assistance has been given by Professor R. DeC. Ward of Harvard University and by Dr. Charles F. Brooks of the Signal Corps School of Meteorology.

Considerable portions of the book are taken verbatim from the pages of the Journal of the Franklin Institute, particularly from papers on the Physics of the Air, by Dr. W. J. Humphreys. The Franklin Institute has also generously provided the electrotypes for 40 illustrations. The Chief of the United States Weather Bureau, Dr. C. F. Marvin, has supplied electrotypes for 31 illustrations and otherwise placed the resources of the Bureau at the disposal of the National Research Council.

On account of the very short time available for the preparation of the book it is unlikely that errors and other defects are lacking. Criticism is invited to the end that a second edition may show improvement and better adaptation to its purpose. Communications may be addressed to Professor Herbert E. Gregory, Yale University, New Haven, Conn.

Division of Geology and Geography,
National Research Council.
Washington, D. C.
September 10, 1918.

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CHAPTER I.

THE ATMOSPHERE.

SOURCES OF METEOROLOGICAL INFORMATION.

THE general means of obtaining information in regard to the atmosphere, and the vertical distribution of meteorological phenomena are shown in Fig. 1. Mountains and other irregularities of the earth's surface make it possible to study the atmosphere minutely and to obtain continuous records of its temperature, pressure and other

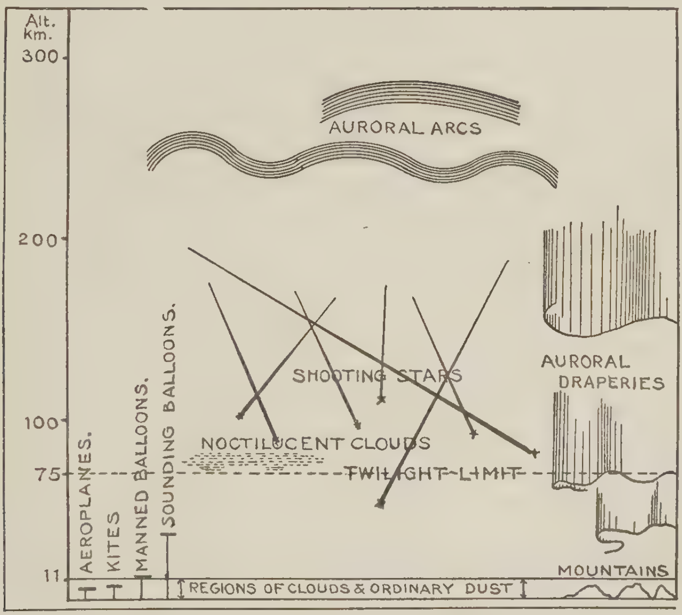

FIGURE 1. Sources of Meteorological Information.

conditions at every elevation from even below sea level (in the regions of the Dead Sea, and Salton Sea, for instance) up to nearly six kilometers above it. Indeed, many automatic records have been obtained at the summit of El Misti, Peru, whose altitude is 5852 meters. Occasionally records have been obtained by this means up to about 7 kilometers, but no higher, as this is the limit to which any

one has ever yet been able to climb. But all such records, whatever the altitude, and however carefully taken, are more or less affected by the surface conditions—they are records of the atmospheric conditions near the surface of the earth. Hence if one would know the conditions of the free atmosphere he must employ some means other than carrying apparatus about on the surface of the earth. One obvious source of information in regard to motion only, and one extensively used, is the drifting of clouds which occur at all levels from the bottom of the atmosphere up to 11 kilometers, or thereabouts, in middle latitudes, and often much higher, 15 kilometers or so, in the tropics, as will be explained later.

There are several methods of determining the elevation, direction of travel and velocity of clouds, but all depend upon simple processes of triangulation. Thus simultaneous observations made with transits on the same spot in a cloud from two stations whose elevations and distance apart are known, obviously furnish all the data necessary for an easy determination of the height of the particular spot in question, while a single subsequent observation by either instrument of this spot, together with the time interval between the first and second observations, gives all the additional data necessary to the determination of its velocity and direction of travel—assuming uniform motion and constancy of elevation. But the direction and velocity of the wind at the time and place of observation is practically all the information about the atmosphere that clouds give, and indeed some clouds, those that hover on or along mountain tops for instance, do not furnish even this. Of course one knows that in the midst of a cloud the air is saturated, or nearly so, and that outside it is unsaturated, but as will be explained later this has very little significance if one has no knowledge of the temperature.

However, there are several other sources of information concerning the free atmosphere; the most fruitful of which is the carrying of self-registering thermometers, barometers, hygrometers, anemometers, etc. by means of:

a. Kites, to over 7 kilometers, the record being 7.26 kilometers. Fig. 2.

b. Aeroplanes, to at least 8 kilometers, with still greater altitudes in prospect.

c. Manned balloons, to about the same height, usually, as aeroplanes, but with a maximum record of roughly 11 kilometers.

d. Sounding balloons, usually to 10 to 20 kilometers, but often higher, the maximum height reported being 35.08 kilometers. Figs. 3 and 4.

In addition to the above, pilot balloons, small balloons without apparatus, and said to have been observed up to a maximum elevation of 39 kilometers, are also used in obtaining, by observations with transits, the direction and velocity of the wind at different elevations on clear days.

The registering apparatus sent aloft by the various methods, *a* to *d*, furnishes reliable information concerning the composition (including humidity), temperature, pressure, direction of motion, and velocity of the air from the surface of the earth up to the greatest elevations attained.

Beyond the reach of the pilot balloon, or, for the present, at elevations greater than 39 kilometers, our information in regard to the atmosphere is limited to such deductions as properly may be drawn from the height to which the sky is illuminated at the end of twilight, as deducted from the angular depression of the sun at that time—roughly 75 kilometers; the paths of shooting stars, rarely if ever seen as high as 200 kilometers; and the phenomena of the auroras, those curious and but partially explained electrical discharges that seldom occur at a lower level than 90 kilometers or higher than 300.

The above are all, or nearly all, the sources of our knowledge of the atmosphere. Up to 35 kilometers above sea level the composition and condition of the atmosphere are comparatively well known, but beyond that level both become increasingly uncertain with elevation.

COMPOSITION OF THE ATMOSPHERE.

If we disregard such obviously foreign things as dust, fog, and cloud, then whatever remains of the atmosphere appears to be ideally homogeneous, and for many purposes it may conveniently be so treated. By the average person, perhaps, as formerly by the ancient Greek philosophers, the atmosphere is supposed to be just what it seems to be—an element in the strictest sense, a thing indivisible into dissimilar parts.

In reality, however, it is not even a single substance, much less a single element, but a mixture of a number of gases and vapors that radically differ from one another in every particular; nor are even the relative percentages of the several distinct constituents at all constant. The story of the chemical conquest of the atmosphere, from the calcination and combustion experiments of the seventeenth and eighteenth centuries that established its complexity down to the refined analyses of the present day that note and account for even the faintest

traces, is full of instruction and inspiration. However, it is practicable to give here only some of the final results.

According to Hann,[1] the chief independent gases that are blended into a dry atmosphere at the surface of the earth, and their respective volume percentages, are as follows:

Element	Nitrogen	Oxygen	Argon	Carbon dioxide	Hydrogen	Neon	Helium
Volume, per cent	78.03	20.99	0.94	0.03	0.01	0.0012	0.0004

In addition to these, krypton and xenon also occur as permanent constituents of the atmosphere. There are also many substances, such as radio-active emanations, the oxides of nitrogen, ozone, and, above all, water vapor, that are found in varying amounts, but of these only water vapor commonly forms an appreciable percentage of the total atmosphere, a percentage that depends chiefly upon temperature in the sense that, for any given pressure, the higher the temperature the greater the possible percentage of water vapor. This relation holds up to the boiling-point of water at the given pressure, when, assuming saturation, there will be nothing but water vapor present, or its percentage will become 100.

Because of this relation of water vapor to temperature its volume percentage decreases from the equator towards the poles, while that of each of the other constituents of the atmosphere correspondingly increases. The annual average values, again quoting from Hann, are:

	Nitrogen	Oxygen	Argon	Water vapor	Carbon dioxide
Equator	75.99	20.44	0.92	2.63	0.02
50° N.	77.32	20.80	0.94	0.92	0.02
70° N.	77.87	20.94	0.94	0.22	0.03

Except for the change in the amount of water vapor, the composition of the surface atmosphere is substantially the same at all parts of the earth. Its composition at different elevations, however, probably differs greatly, but a discussion of this subject involves the use of more mathematics than it is convenient to give briefly. Hence only the general facts will be presented.

Table I, computed by the aid of the necessary hypsometric equations, and Fig. 5, drawn in accordance with this table, give the approximate composition and barometric pressure of the atmosphere at various levels. The assumptions upon which they are based are in close agreement with the average conditions of middle latitudes, and are as follows:

[1] Lehrbuch der Meteorologie, 3d ed., p. 5.

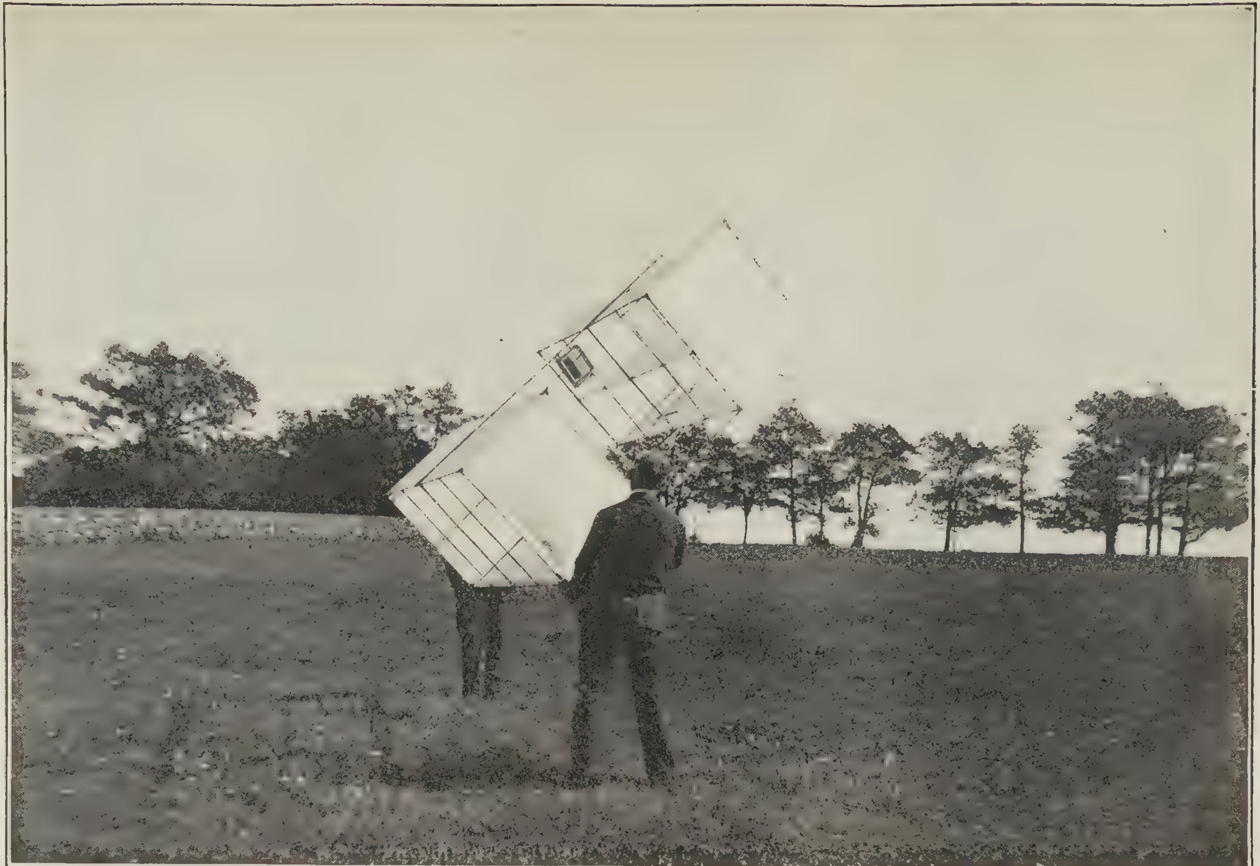

FIGURE 2. Launching meteorological kite, Mount Weather, Va.

Figure 3. Sounding Balloon.

Figure 4. Sounding Balloons.

TABLE I.

Percentage Distribution of Gases in the Atmosphere.

Height in kilometers	Gases							
	Argon	Nitrogen	Water vapor	Oxygen	Carbon dioxide	Hydrogen	Helium	Total pressure in millimeters
140	....	0.01	....	.....	....	99.15	0.84	0.0040
130	....	0.04	....	.....	....	99.00	0.96	0.0046
120	....	0.19	....	.....	....	98.74	1.07	0.0052
110	....	0.67	0.02	0.02	....	98.10	1.19	0.0059
100	....	2.95	0.05	0.11	....	95.58	1.31	0.0067
90	....	9.78	0.10	0.49	....	88.28	1.35	0.0081
80	....	32.18	0.17	1.85	....	64.70	1.10	0.0123
70	0.03	61.83	0.20	4.72	....	32.61	0.61	0.0274
60	0.03	81.22	0.15	7.69	....	10.68	0.23	0.0935
50	0.12	86.78	0.10	10.17	....	2.76	0.07	0.403
40	0.22	86.42	0.06	12.61	....	0.67	0.02	1.84
30	0.35	84.26	0.03	15.18	0.01	0.16	0.01	8.63
20	0.59	81.24	0.02	18.10	0.01	0.04	....	40.99
15	0.77	79.52	0.01	19.66	0.02	0.02	....	89.66
11	0.94	78.02	0.01	20.99	0.03	0.01	....	168.00
5	0.94	77.89	0.18	20.95	0.03	0.01	....	405.
0	0.93	77.08	1.20	20.75	0.03	0.01	....	760.

1. That at the surface of the earth the principal gases of the atmosphere and their respective volume percentages in dry air are:

Nitrogen	78.03	Neon	0.0012
Oxygen	20.99	Helium	0.0004
Argon	0.94	Carbon dioxide	0.03
Hydrogen	0.01		

2. That at the surface of the earth water vapor supplies 1.2 per cent. of the total number of gas molecules present.

3. That the absolute humidity rapidly decreases, under the influence of lower temperatures, with increase of elevation, to a negligible amount at or below the level of 10 kilometers.

4. That the temperature decreases uniformly at the rate of 6° C. per kilometer from 11° C. at sea level to −55° C. at an elevation of 11 kilometers.

5. That beyond 11 kilometers above sea level the temperature remains constant at −55° C.

6. That up to the level of 11 kilometers the relative percentages of the several gases, excepting water vapor, remain constant—a result of mixing by air currents (vertical convection).

7. That above 11 kilometers, where the temperature changes but

little with elevation, and where vertical convection, therefore, is practically absent, the several gases are distributed according to their respective molecular weights.

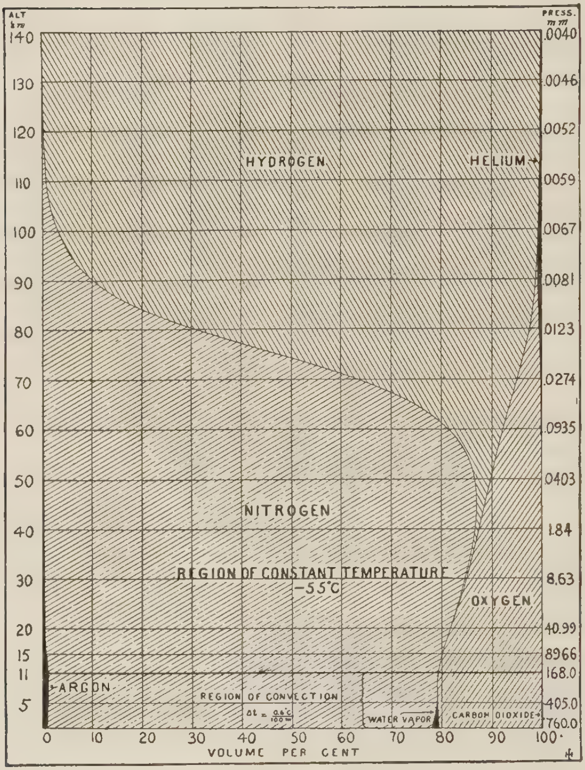

FIGURE 5.

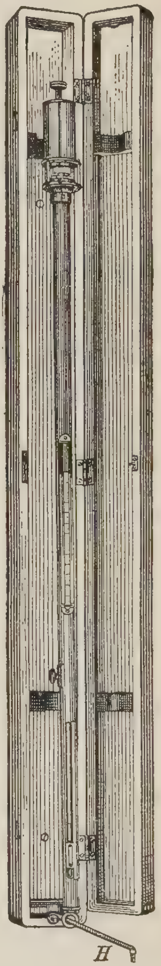

FIGURE 6.

FIGURE 5. Composition of the Atmosphere at Different Levels.
FIGURE 6. Mercurial Barometer in box.

A number of atmospheric gases—neon, krypton, xenon, ozone, etc.—are omitted both from Table I and from its accompanying figure. This is because all these occur—in the lower atmosphere, at any rate—in quantities too small for graphical illustration in the same diagram with the principal gases and to the same scale.

In using this diagram it should be distinctly remembered that it is supported by direct experimental observations only from the surface of the earth up to a level of about 30 kilometers, and that, while the extrapolated values are based upon apparently sound logic and not mere surmises, they necessarily become less and less certain with increase of elevation.

The table and the figure bring out a few points not generally realized. One of these is the fact that the total amount of argon in the atmosphere is much greater than the average total amount of water vapor. Another is the surprisingly small amount of water vapor, especially in view of the wonderful things it does, and of its vital importance to life of every kind. There may also be a little surprise that, according to calculation, the percentage of water vapor reaches a certain maximum at an elevation of 70 to 80 kilometers, where it is, roughly, twenty-fold what it is at, say, 11 kilometers. This, however, does not mean that the total amount of water vapor increases with elevation, but that it decreases less rapidly than do the heavier constituents, and more rapidly than the two lighter ones, hydrogen and helium.

CHAPTER II.

MEASUREMENT OF THE METEOROLOGICAL ELEMENTS.

GENERAL CONSIDERATIONS.

The more important meteorological elements, easily measureable by means of simple apparatus, are, the pressure of the atmosphere, temperature of the air, direction and velocity of the wind, humidity, rainfall and cloudiness. Much more difficult to obtain and requiring special apparatus in the hands of trained observers, are measurements of evaporation, solar and terrestrial radiation, atmospheric electricity, and the vertical distribution of the meteorological elements.

Before undertaking any work, however elementary in character, even if the equipment required is very simple, students should familiarize themselves with the best methods of taking and recording observations and the construction and proper use of the apparatus to be employed. Much useful information concerning these subjects is easily accessible, in text-books of meteorology,[1] catalogues of the best instrument-makers,[2] and particularly in circulars of instruction published by the various weather services.[3] Literature of this kind studied in connection with practical work with instruments ordinarily will be sufficient for the needs of an intelligent observer, but proficiency will be acquired much more rapidly if the student can avail himself of competent instruction or opportunities of studying the work of a first-class meteorological station.

Nearly all meteorological studies are comparative, and the most frequent use of data collected at any station is in comparison with similar data obtained at other stations; consequently the value of observations or records is greatest when they are made by trained observers employing apparatus of standard quality according to methods approved by the best authorities. But it must not be inferred that costly or unusual instruments are always necessary. Investiga-

[1] See bibliography at end of book, Appendix I.

[2] Catalogues of Jules Richard, Paris; Negretti & Zambra, London; Henry J. Green, New York; Julien P. Friez, Baltimore; Taylor Bros., Rochester, N. Y.

[3] See bibliography at end of book, Appendix I.

tions of the highest degree of excellence have been accomplished by means of simple, inexpensive, or even inferior instruments, whose errors or defects had been determined by comparisons with standards. Instruments of inferior materials or workmanship usually are neither so durable nor are their errors so constant as those of a standard quality, and require more frequent examinations for defects.

DIRECT-READING INSTRUMENTS.[4]

Temperature.—In meteorology, the primary use of the thermometer is for the purpose of determining the temperature of the free air; that is, the temperature of the air uninfluenced by the direct rays of the sun, by heat radiated by or conducted from objects exposed to direct sunlight, or by artificial heat. The instrument adapted to the greatest variety of uses is the mercurial thermometer, having a cylindrical bulb and graduated upon its stem. For ordinary purposes thermometers are graduated to single degrees or half-degrees and can be read to tenths of degrees by estimate; the errors should not exceed $\pm 0.2°$ C. and should be uniform throughout the scale.

The mercurial thermometer becomes useless when the temperature falls to the freezing-point of mercury, ($-39°$), and at this point must be replaced by thermometers filled with alcohol, or (if extremely low temperatures are to be measured), toluene. Alcohol-filled thermometers are also used in registering minimum temperatures. All thermometers of this kind are less sensitive than mercurial thermometers, have larger errors and are more liable to get out of order.

Pressure.—The pressure of the atmosphere may be determined very easily by measuring the length of a column of liquid such as mercury, glycerine, water, etc., that it will support or the extent to which it will compress an elastic body. Since, at sea-level the pressure of the air is 1033.3 grams per square centimeter approximately and the weight of mercury at 0° C. is 13.596 grams per cubic centimeter, the average height of a column of mercury whose weight is equal to that of a column of the atmosphere of the same cross-section is 76 centimeters. Atmospheric pressures usually are expressed as direct measures, in standard units, of the length of the vertical mercurial column which they balance. Recently, however, international authority has recom-

[4] Most of the instruments described in this chapter are in use by the United States Weather Bureau. It is very desirable to arrange with the local observer for permission to examine the instruments in his office.

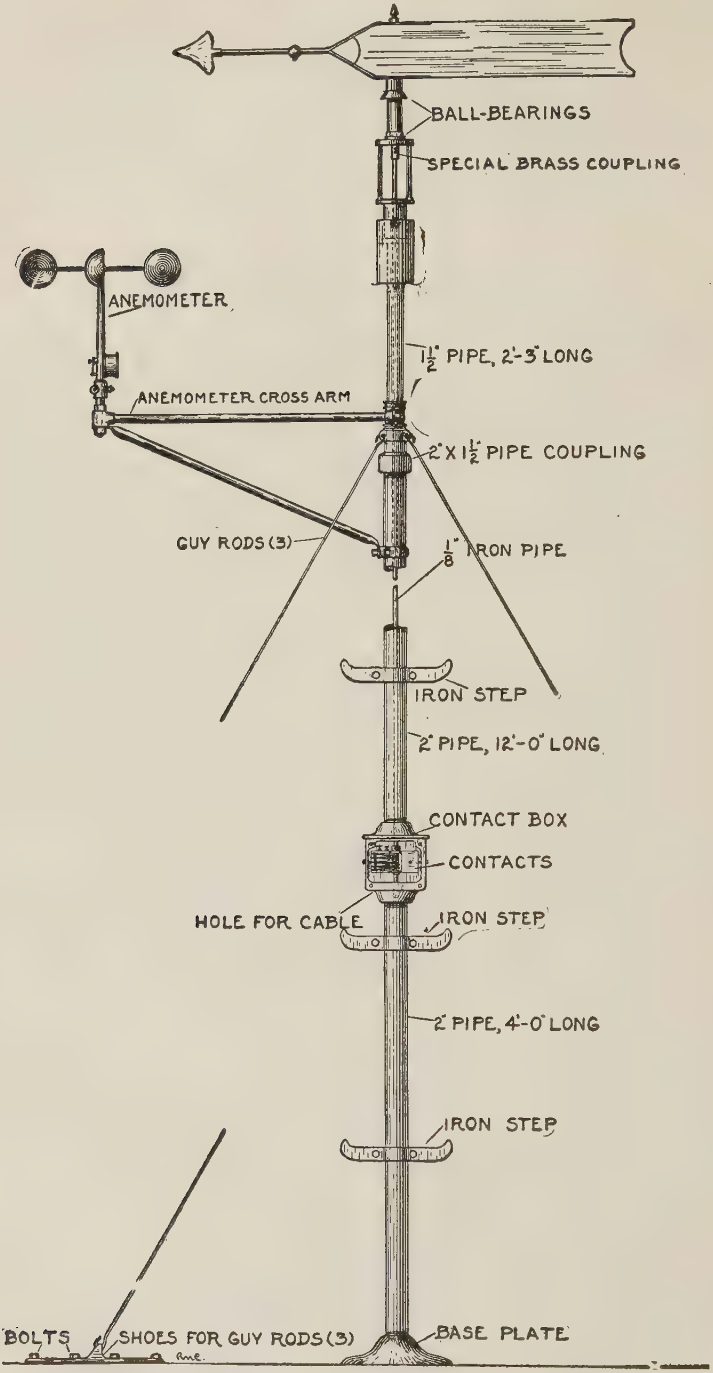

FIGURE 7. Anemoscope and Anemometer upon standard support.

mended that pressures be measured in units of the C. G. S. system,[5] in which the standard atmosphere is 1000 millibars, approximately equal to a barometric reading of 750 millimeters, the average value at a height of about 106 meters above sea-level. The pressure in millibars at any place is a direct percentage of the standard atmosphere and comparisons are very easily made. Another important advantage of the millibar is that since it is smaller than the millimeter (1 mb. = 0.75 mm.), barometric readings to 0.1 mb. (0.075 mm.) which are sufficiently refined for most purposes, can be made without the use of a vernier.

Of the many different forms of the mercurial barometer the one devised by Fortin and having a cistern with a fixed zero has come into general use in meteorology. The pattern of Fortin barometer employed in the work of the United States Weather Bureau is shown in Figure 6.

The height of the barometric column is influenced by temperature, gravity and differences of elevation, for all of which appropriate corrections must be made before the readings can be used in meteorological studies.

While for observations of the highest degree of accuracy and uniformity the mercurial barometer is indispensable, it is a delicate instrument and cannot be moved without more or less danger of breaking it or impairing the vacuum. To meet the demand for a compact portable instrument, various forms of the so-called aneroid barometer have been devised, and for certain purposes, such as determining the pressure during ascensions of balloons, kites or aeroplanes, no other form of instrument can be used. These instruments are very sensitive, but at best they are not uniformly accurate and their errors are so variable that frequent comparisons with mercurial barometers are necessary in order to obtain satisfactory results.

Direction of the Wind.—When it is desired to ascertain the direction of the wind only to four or eight principal points, direct observations of a well-exposed vane or anemoscope[6] usually will be sufficient. Greater accuracy or convenience is gained when the axis of the vane extends into a room below and indicates changes of direction upon a dial, or the instrument can be made to record electrically at any place desired, as does the United States Weather Bureau anemoscope. This instrument, mounted upon a standard support also carrying an anemometer, is shown in Figure 7.

[5] A system of units commonly used in physics, based upon the centimeter as the unit of length, the gram as the unit of weight or mass, and the second as the unit of time.

[6] The terms *vane* and *anemoscope* are synonymous.

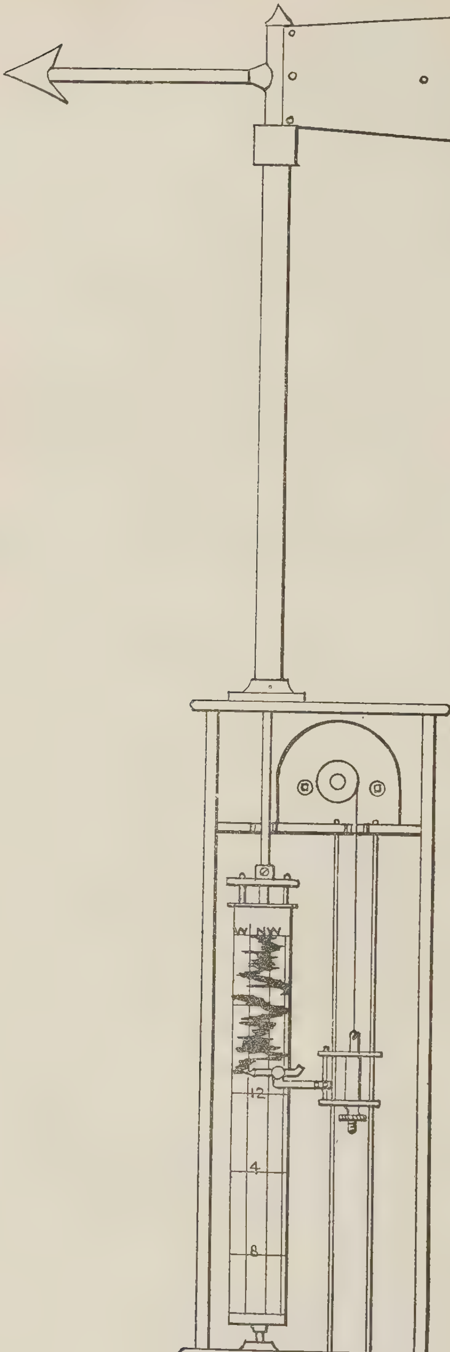

FIGURE 8. Portable Draper Anemoscope.

When it is desired to study minute changes of direction a recording anemoscope of the Draper pattern should be employed. A portable form of this instrument, useful either at permanent or temporary stations, is shown in Figure 8.

Movement or Velocity of the Wind.—Two classes of instruments

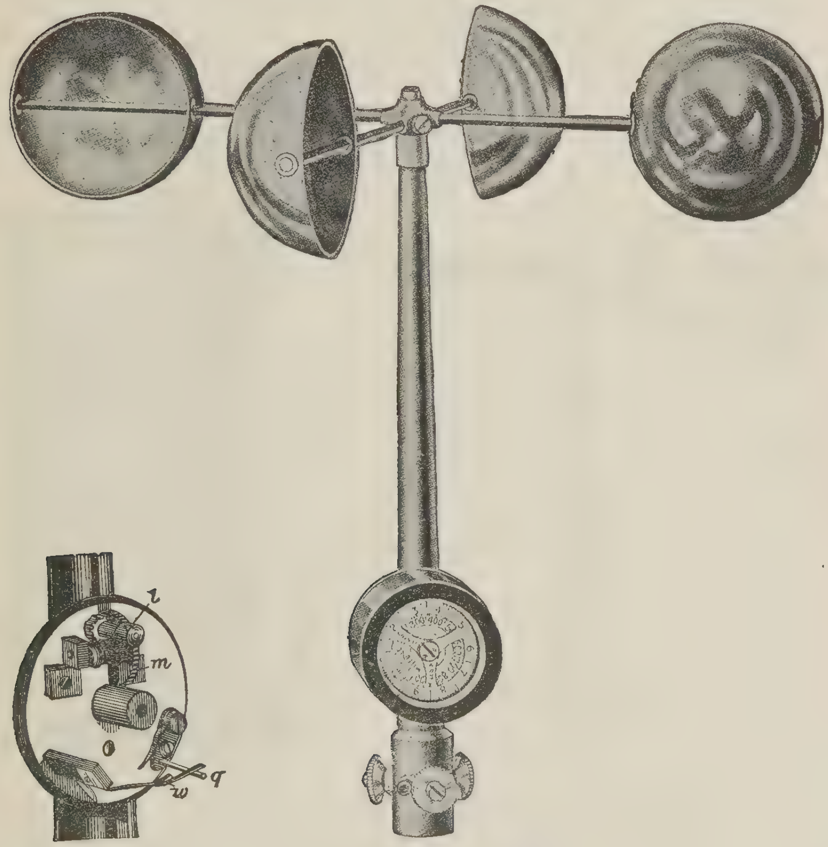

FIGURE 9. Robinson's Anemometer, Weather Bureau pattern.

are employed in measuring the energy of the wind, velocity or "rotation" anemometers and pressure anemometers. In the former are instruments of the Robinson and windmill or "screw" patterns, in which the movement and velocity are determined from the rotations of systems of cups or fans moved by the wind. In the latter are the various pressure plates and instruments of the Lind, Dines, and

Venturi type, which measure the direct pressure of the wind upon surfaces of known area or upon the liquid in a manometer (pressure-gauge). Excellent instruments of both classes are in use and yielding satisfactory results.

For the largest variety of uses, and where it is desired to measure mean or average velocities, the Robinson anemometer is probably

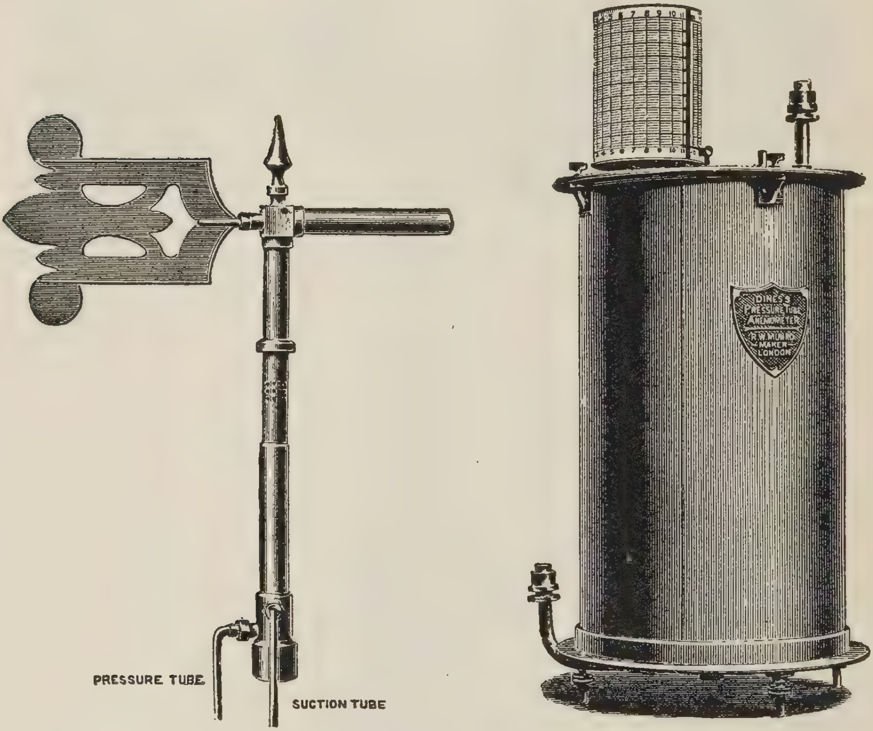

FIGURE 10. Dines' Pressure-tube Anemograph.

superior to all other instruments, and the pattern adapted by the Weather Bureau is well suited to different conditions and uses either as a direct-reading or recording instrument. Details of this instrument are shown in Figure 9, and the customary method of exposure in Figure 7.

For indicating rapid variations of velocity and for certain special uses, anemometers of the "windmill" pattern or those of the pressure type are superior to the Robinson instrument. The most extensively-

used windmill anemometer is Richards' anemo-cinemometer in which a light six-bladed screw or fan makes one rotation while the wind moves one meter. . Of pressure instruments, probably the most satisfactory is Dines's tube anemometer (Figure 10), in which the pressure of the wind in a tube causes a change of level of the liquid in a manometer or operates a float bearing an index or recording pen.

Pressure of the Wind.—Within a range indicated approximately by velocities of 3 and 50 meters per second the pressure of the wind varies nearly as the square of the velocity. In any instance the actual pressure also depends upon the character and dimensions of the surface and the density of the air which, in turn, is a function of the barometric pressure and the temperature of the air. At sea-level, under ordinary conditions, wind-pressures may be determined with fair accuracy by the formula:

$$P = 0.0735SV^2$$

in which P is the pressure in kilograms per square meter, of surface exposed normally, S the surface in square meters, V the velocity in meters per second, and 0.0735 a factor determined by experiment.

Moisture or Humidity of the Air.—For nearly all the purposes of meteorology the moisture of the air, whether expressed as the dew-point, vapor pressure, or absolute or relative humidity, may be determined most conveniently by means of observations of the psychrometer or dry-bulb and wet-bulb thermometers. Of the many different forms of this instrument the sling-psychrometer (Figure 11) is probably the best for general use.

When the temperature falls below the freezing-point of water, observations of the psychrometer require much greater care than is the case with higher temperatures, and when very low temperatures prevail other forms of hygrometers must be used. The best of these are the hair-hygrometers, of which the best-known patterns are those designed by Koppe, Lambrecht (Figure 12), and Richard (Figure 13). The chief defect of hair-hygrometers is the uncertain change of zero, which necessitates frequent comparisons with psychrometers or other standards.[7]

Precipitation.—The amount of rain is the depth to which it would accumulate on a horizontal surface if the water remained in the place where it fell. Any vessel with vertical sides can be used as a rain-gauge, but greater accuracy is obtained if the amount caught in a

[7] Annals, Harvard College Observatory, LVIII, 2, 1906.

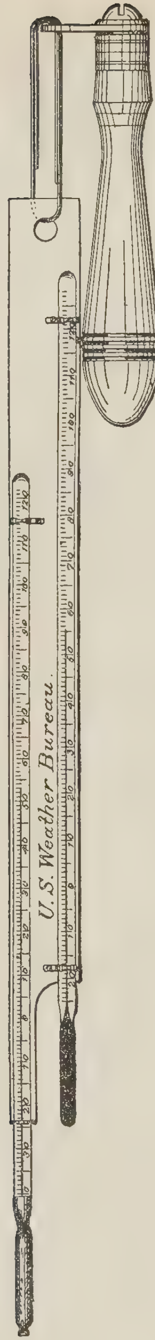

FIGURE 11. Sling-psychrometer.

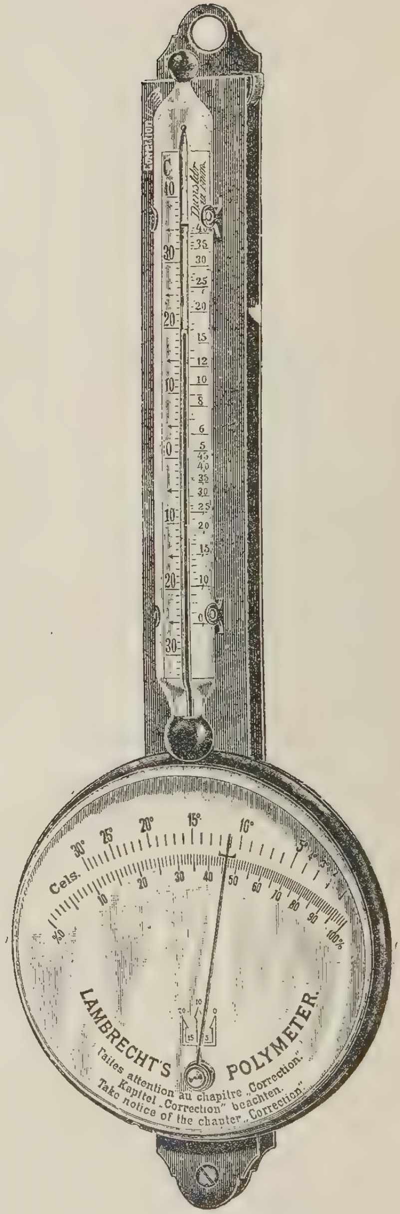

FIGURE 12. Lambrecht's Hygrometer.

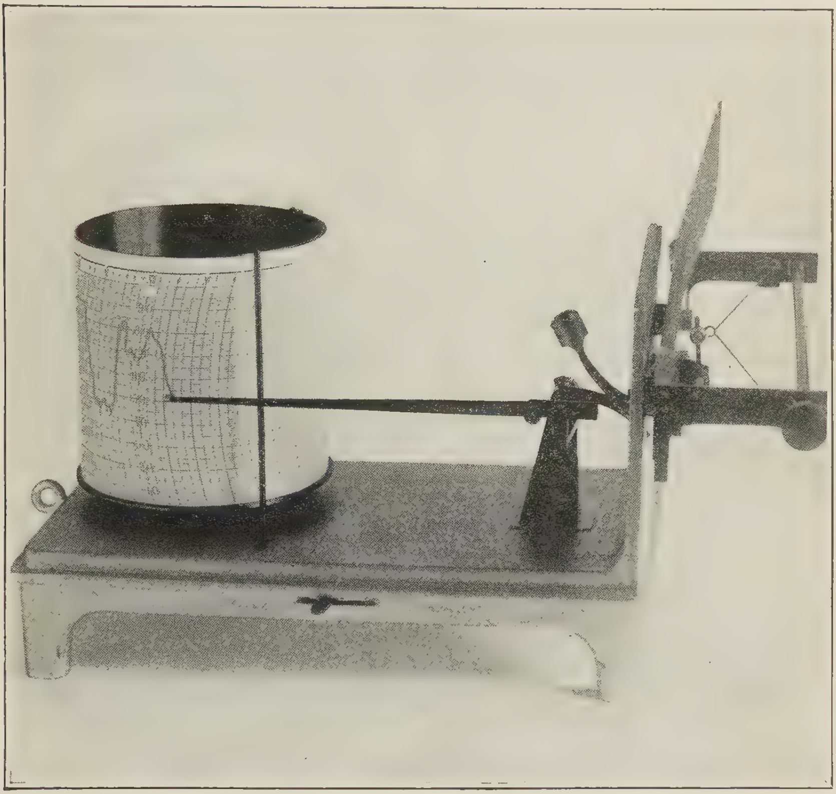

Figure 13. Richard's Hygrograph.

receiver of any size is measured in a much smaller tube, for instance, one having one-tenth the cross-section of the receiver. An excellent instrument of this kind is the United States Weather Bureau raingauge, shown in Figure 14.

In the measurement of snow, both the actual depth and the amount of water the snow contains are important. The water-content is never constant but under different conditions varies from about five to more

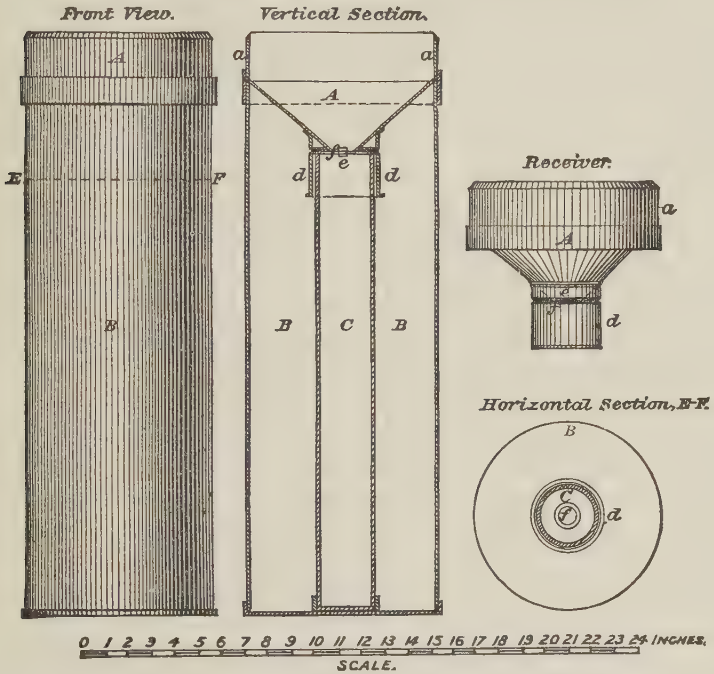

FIGURE 14. Rain-gauge.

than 50 per cent. Snow, being much lighter than rain, is much more difficult of measurement for the reason that the amount caught in gauges, even those of special design, is seriously affected by wind. The most accurate method of measurement is to cut a section of known dimensions from a layer of snow on the ground (taking care of course to select a place when the covering is of average depth) and determine the water-content, either by weighing the dry snow on a suitable scale or melting it and measuring the water as so much rain. In regions where the snow accumulates to great depths, special instruments are

necessary, of which the Church sampler is probably the best.[8] This instrument is made of steel tubing, of small size, in sections united by a special coupling so that one section or the entire number can be used according to the depth to be measured. The depth is measured by a scale ruled on the tube and the water-content of the sample is found by weighing the tube and its contents on a scale compensated for the weight of the empty tube.

Evaporation.—The phenomenon of evaporation has received much serious attention within recent years. Consideration of the results obtained by numerous investigators indicates that practically every research in this field requires its own special equipment and methods of work, these depending, for example, upon whether soils, vegetation or open surfaces of water are to be studied. At the present stage of our knowledge, all measures must be considered as relative, for even in the instance of evaporation-pans extensively used, the results depend very greatly upon local conditions of temperature, moisture, wind, etc., and records at different places are not strictly comparable.

The Direction and Movements of Clouds.—As in the instance of the winds, the direction of motion of clouds, under ordinary or average conditions, can be determined to principal points of the compass, without apparatus. But in the instance of slow-moving clouds, or if very accurate observations are desired, a nephoscope will be necessary. A simple nephoscope may be made of a circular mirror by securing to its rim a scale ruled in points of the compass or (preferably) in degrees of azimuth numbered from 0° south through 90° (west) around to south again, and cutting a small cross-mark in the center for use as a zero-point. When in use the instrument is placed upon a table or stand with its zero or south point toward the north so that a line through the center of the mirror connecting the north and south points is parallel to the meridian of the place of observation. Observations of the direction of clouds are made by noting the movement of their reflected images from the center to the rim of the mirror; the reading of the scale at the point *toward* which the images move is the direction or azimuth *from* which the clouds are moving.

The relative velocity, from which the actual velocity of clouds may be computed, is the distance or space traversed by the cloud image in a given time when the observer's eye is held at a definite height above the mirror. One very simple rule is to note the distance, in millimeters, traveled in one minute by the cloud-image while the eye

[8] Scientific American Supplement, September 7, 1912.

is 16⅔ centimeters above the surface of the mirror; this distance or relative velocity multiplied by the height of the cloud in meters and divided by 10,000 is equal to the actual velocity in meters per second.[9] In practically all cases, by dropping the ciphers, the computation can be performed mentally, as for example, with a relative velocity of 125 mm. and height of cloud of 1500 meters, we have $125 \times 15 = 18.7$ meters per second. This method of obtaining actual velocities is accurate only for clouds in the cumulus level or lower, or when the height is known, but will give approximate values for all levels when it is employed by well-trained observers. According to researches by Clayton, Vettin and others, the relation of cloud-form to height is constant and universal, and the primary forms are most frequently formed in five different levels above the earth's surface.[10] In the North-eastern United States these levels, indicated by the prevailing cloud-form, and their height in meters are: Stratus level, 500; Cumulus level, 1600; Alto-Cumulus level, 3800; Cirro-Cumulus level, 6600; Cirrus level, 8900. (See Chapter VI.) Hence, if the relative velocity is measured of any cloud easily assignable to one of the levels, the actual velocity can be determined with fair accuracy, particularly if the time of year and the conditions prevailing at the time of observation are taken into consideration. Obviously the method becomes uncertain when applied to ill-defined or intermediate forms, but even here the skilled observer will obtain results of value.

As suggested by Espy, the heights of cumulus clouds can be computed from observations of the dew-point (Chapter V). In Clayton's very simple formula the height in meters is obtained by dividing the difference between the air-temperature and the dew-point by the average temperature-gradient (or decrease of temperature with increase of height above sea-level) existing between the earth's surface and the point where condensation begins, applying a correction for expansion of the air.[11] The temperature-gradient is approximately 0.97° C. for each 100 meters of elevation. On account of expansion of the ascending air the dew-point is lowered 0.2° C. for each 100 meters, consequently, the factor used in computing heights of clouds becomes 0°.77. For example, if the temperature is 25° and the dew-point 10° we have $25° — 10° = 15°$; $1500 \div 0.77 = 1949$ meters, or the height at which the ascending moisture condenses into clouds.

[9] Annals, Harvard College Observatory, XLII, 2, pp. 202-203.
[10] Annals, Harvard College Observatory, XXX, 4, pp. 337-342.
[11] Annals, Harvard College Observatory, LXVIII, 1.

SELF-RECORDING APPARATUS.

A continuous record by means of self-recording instruments can be obtained of almost any meteorological element and when checked by comparisons with readings of standard instruments becomes the most useful of all records.

Pressure.—For most uses, continuous records of pressure from aneroid barographs of the Richard pattern, Figure 15, are sufficient. When greater precision is desired mercurial barographs should be employed. The instruments designed by Sprung,[12] Marvin[13] and Draper[14] are probably the best known of this class.

Temperature.—Richard's thermograph, Figure 16, in which the temperature is measured by means of a Bourdon pressure-tube filled with liquid, is probably the most satisfactory instrument for general use. Many different scales and degrees of sensitiveness are possible with this instrument. For special researches, and where great sensitiveness is desired, electrical-resistance instruments have been employed with success, but are much more difficult of operation.

Humidity.—Absorption hygrometers, in which the relative humidity is determined by the changes of length of a strand of human hair, can easily be so constructed as to give a continuous record, as in the instance of Richard's hygrograph shown in Figure 13. These instruments are very sensitive, but require more careful supervision than thermographs or barographs of the same general types.

Wind.—For recording the direction and velocity the equipment best adapted to a variety of uses and conditions is the Weather Bureau anemoscope and anemometer already described, connected electrically with the meteorograph or "quadruple register" where the direction to eight principal points, movement of the wind, duration of sunshine and rate and amount of rain are recorded upon one sheet of paper moved by clock-work. Electrically-recording instruments are usually to be preferred where the recording mechanisms must be placed at a distance from the parts exposed to the wind.

The Dines anemometer, Figure 10, and the Richard anemo-cinemograph possess important advantages over other instruments, in that velocities are read directly from the records without computation, and variations in velocity are more obvious to the eye than is the case with records where a mark is made at the end of each kilometer, etc.,

[12] Report of Chief Signal Officer, 1887, Part 2.

[13] Monthly Weather Review, September, 1908.

[14] American Meteorological Journal, December, 1884.

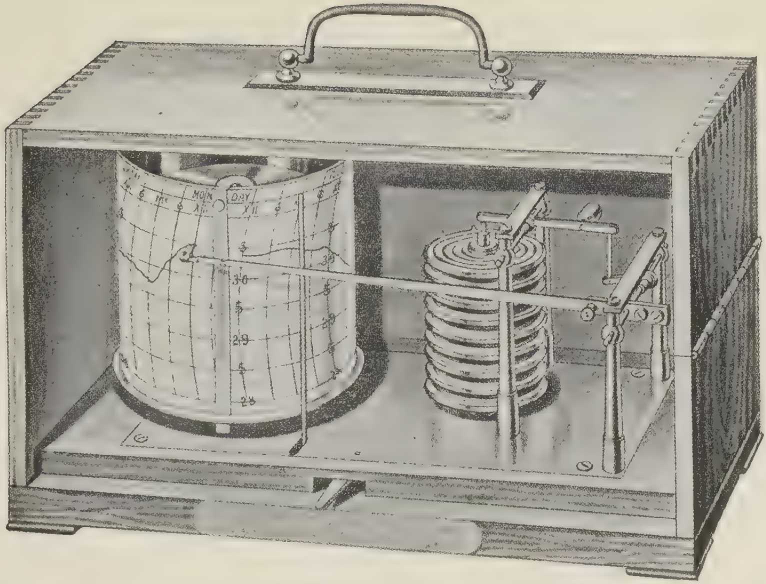

FIGURE 15. Richard's Aneroid Barograph.

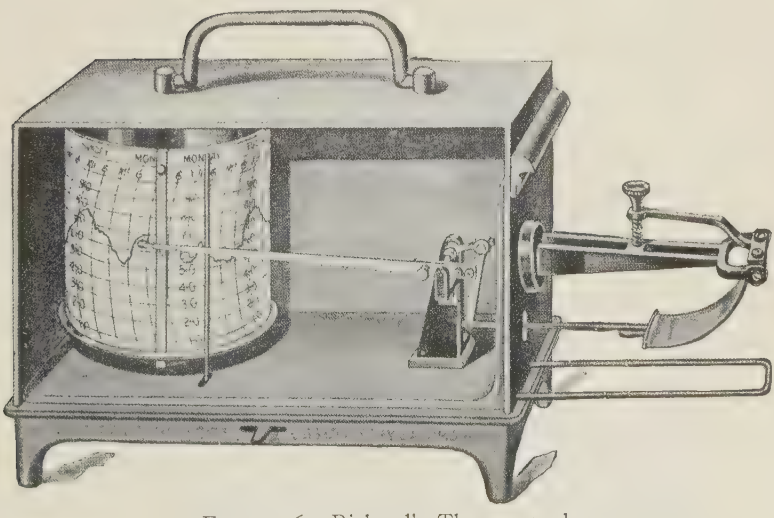

FIGURE 16. Richard's Thermograph.

traveled by the wind. Also, these instruments are better adapted to recording details of gusts and extreme variations than are the Robinson anemometers.

Precipitation.—The tipping-bucket raingauge, Figure 17, is prob-

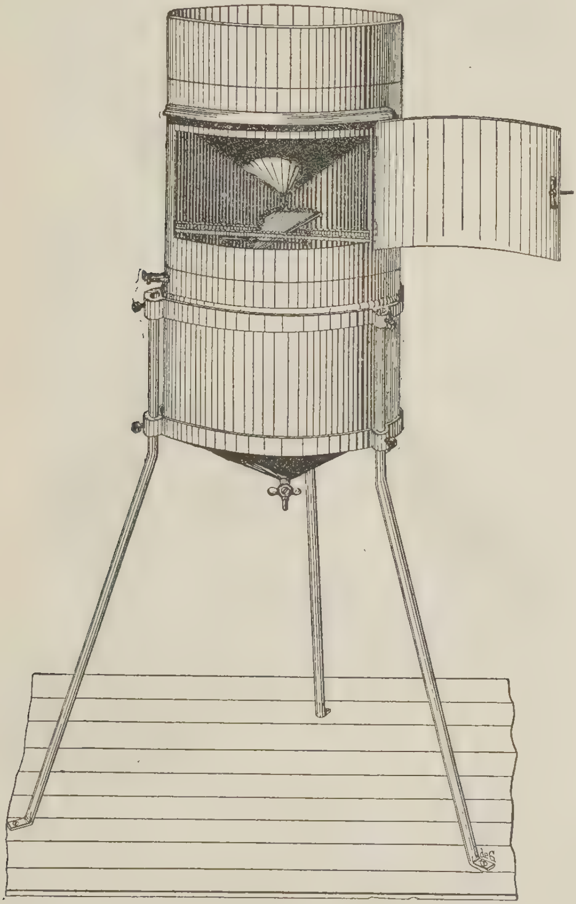

FIGURE 17. Tipping-bucket Rain-gauge.

ably the most convenient for many purposes where it is important that the record should be visible while rain is falling, but the amount recorded by instruments of this class varies with the rate of rainfall, and where the precipitation is chiefly in the form of snow, devices for melting the snow become necessary. For recording rain, snow,

hail or mixed precipitation, weighing gauges are best. Successful examples of this kind are the Marvin electrically-recording gauge, the Richard, Draper, and Fergusson instruments, the last three being self-contained and adapted for use as seasonal instruments.

Sunshine and Cloudiness.—The thermometric sunshine recorder, Figure 18, recording electrically upon the quadruple register already referred to, is employed at all stations of the United States Weather

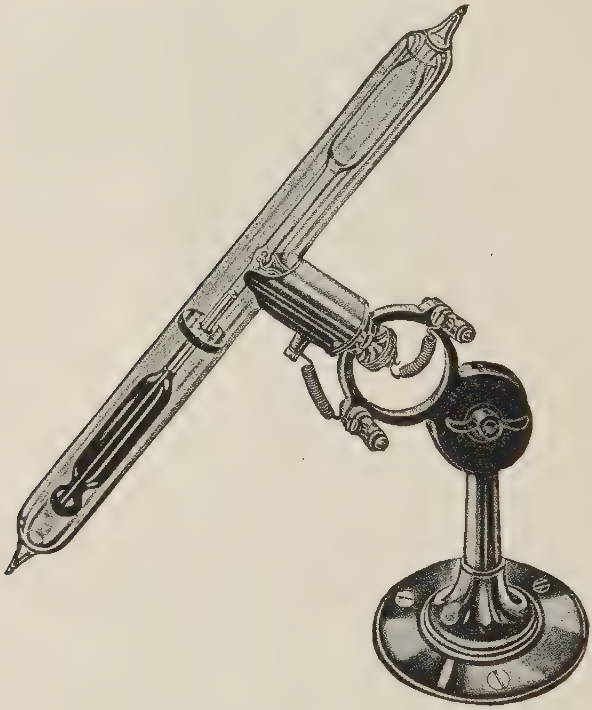

FIGURE 18. Thermometric Sunshine Recorder.

Bureau. Where electrically-recording apparatus is not available, the photographic instrument shown in Figure 19 can be substituted for the thermometric. The chief defect of the latter instrument is that the blue-print paper used for record-sheets is not always of uniform quality and sensitiveness, and unless allowance is made for this the records are not strictly comparable. Only freshly-sensitized paper should be used. The Campbell-Stokes recorder, employed as a standard in Europe, consists of a spherical lens so mounted in a circular frame that a strip of cardboard, also held by the frame in the focus of the lens, is charred by the concentrated rays of the sun. The duration,

and under certain conditions the intensity of bright sunshine, are determined by the length of the space charred. Records from this instrument probably are more uniform and therefore more nearly comparable than are those of other sunshine recorders.

The only instrument suitable for recording the cloudiness at night is the Pickering pole-star recorder, which is a camera of long focus, pointed continuously toward the polar-star.[15] The shutter is kept open during the entire night and the apparent motion of the star about the true pole is photographed in the form of an arc of a circle. On clear nights the star-trail is continuous and on nights more or less cloudy

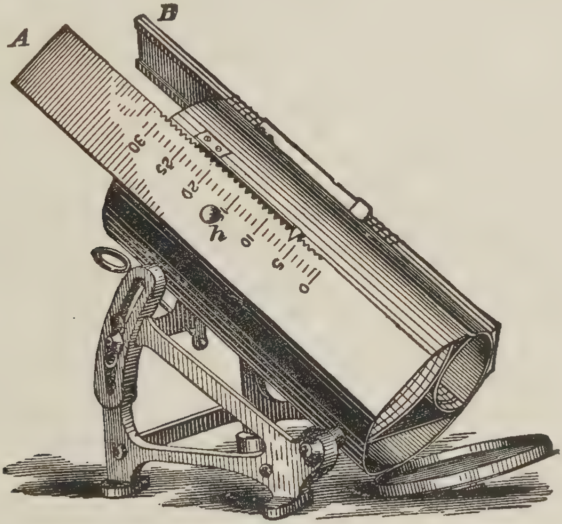

FIGURE 19. Photographic Sunshine Recorder.

it is broken wherever the star is obscured for an appreciable time by the clouds. Records from such instruments are very much to be desired, for the reason that at the present time the few instruments in operation are in Eastern states and little is known concerning the variations of cloudiness at night.

EQUIPMENT FOR INVESTIGATIONS OF THE UPPER ATMOSPHERE.

With the possible exception of measures of clouds, the more important advances in the study of the upper atmosphere (a branch of meteorology to which the name "aerology" has been given) began

[15] Quarterly Journal, Royal Meteorological Society, October, 1905.

with the use of light self-recording instruments, carried by sounding balloons (1893) and kites (1894). While much excellent equipment of all kinds has been suggested and employed, general experience so far has been insufficient to develop standard forms of instruments; each institution or organization usually having prepared its own.

Heights and Velocities of Clouds.—These data are obtained almost entirely by triangulation, by means of simultaneous observations at two stations a known distance apart, measurements of negatives from photogrammeters (theodolites equipped with cameras), soundings by kites or balloons or from observation of the shadows of clouds or the illuminated under surfaces of clouds over well-lighted cities. In simple triangulation, two stations are equipped with transits having sighting tubes instead of telescopes, and connected by telephone so that the observers can identify the clouds to be measured. Synchronism of observation is best obtained by securing to the transmitter of one telephone a telegraph sounder connected with a clock beating seconds, so that both observers may hear the beats.[16] The photographic methods are much more difficult and a description thereof seems unnecessary here. The last two methods, except in rare instances of dense or well-defined upper clouds, are applicable only to lower clouds, but are very easy, requiring only a simple alt-azimuth (an instrument for measuring altitudes and azimuths) or protractor for measuring altitudes.[17] A further application of the last method would be to illuminate clouds at night by means of a powerful search-light, which, under favorable conditions might be used in the measurement of clouds as high as the alto-cumulus level.

Kites and Balloons.—The most important method of securing information concerning the higher atmosphere is that in which light recording instruments are carried aloft by kites, balloons or aeroplanes. A very important advantage possessed by kites is the ease with which records at definite heights may be secured. The Hargrave kite, in its various modifications, has come into general use in aerology, and in Figure 2 is shown the pattern employed by the Weather Bureau, with the meteorograph or recording apparatus secured between the cells. The meteorograph records atmospheric pressure (from which the height is computed), temperature, humidity and wind-velocity; also, the direction of the wind is determined from the position of the kite, observed with a transit or theodolite. The kites are flown by steel music wire controlled by a special power reel or windlass,

[16] Annals, Harvard College Observatory, XLII, 2, p. 194.
[17] Annals, Harvard College Observatory, XXX, 3, p. 207.

equipped with devices for indicating or recording the length of line in use and the tension or pull exerted by the kites. From observations of the length of the line and the angular altitude of the kites, the vertical or linear height is easily obtained. The greatest elevation reached by kites, so far, is 7260 meters.

Captive balloons afford the same class of data as do kites, but are so sensitive to variable winds that their use is comparatively limited and the heights attained are lower. The best method of using such balloons is to reel out the line so rapidly that the balloon rises as a free balloon until the weight of line equals the lift, then drawing it to earth. A very sensitive recording instrument is necessary, and usually the wind-velocities recorded are uncertain.

By means of sounding balloons (Figures 3 and 4), particularly the small expanding balloons devised by Assmann, recording instruments have been carried to a maximum height of 35000 meters.[18] Temperature and pressure, from which the height is computed, and occasionally humidity, are recorded and the direction and velocity of the wind can be determined from observations of the balloon as it rises or falls.

Pilot-balloons, exactly similar to but much smaller than sounding balloons, and carrying no instruments, are also used to measure the direction and movement of the upper air. The exceptional height of 39000 meters has been attained by balloons of this kind. The heights, etc., are determined by triangulation.

ACCURACY OF INSTRUMENTS AND METHODS OF EXPOSURE.

Barometers and Barographs.—Readings by good observers of barometers of the ordinary "station" pattern should agree within 0.06 mm., and the differences between readings of good barometers should not exceed 0.13 mm. Standard barometers, having tubes 15 mm. in diameter, or larger usually, can be depended upon within 0.04 mm.

Records of aneroid barographs in good condition should be accurate within 0.6 mm., and the errors of mercurial barographs of the Sprung, Marvin and Draper patterns ordinarily should not exceed 0.2 mm.

Concerning the exposure of barometers and barographs, the only requirement of importance is that these instruments be kept in a well-lighted room where the temperature is uniform.

Thermometers and Hygrometers.—The errors of good thermom-

[18] Annals, Harvard College Observatory, LXVIII, 1.

eters, even those filled with alcohol, are small, usually not exceeding a few tenths of a degree; and are easily corrected by comparisons with standards. Such instrumntal errors are far less important than those due to faulty exposure.

The accuracy of observations of humidity, particularly at low temperatures, depends very greatly upon the ventilation of the instruments. The greatest accuracy is attained when both psychrometers and hygrographs are continuously ventilated, for often the larger differences found between standard and recording instruments are due to actual local differences of condition in the instrument shelters, or differences in sensitiveness, particularly if there is little or no wind. During calms, in practically all closed louvred shelters, such as the

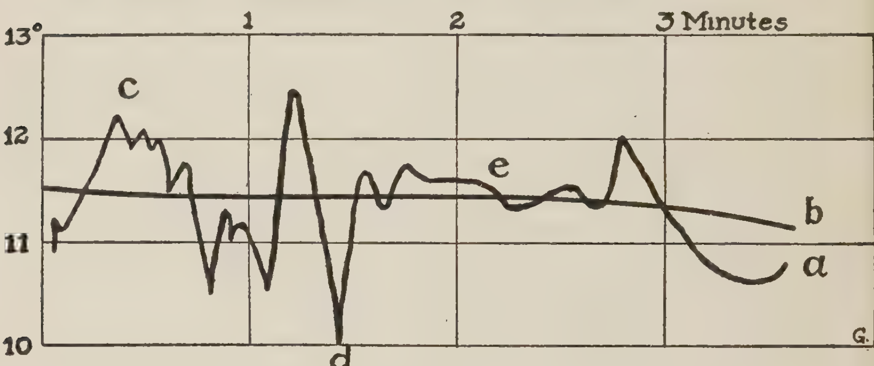

FIGURE 20. Comparative Sensitiveness of Instruments.

cage-like thermometer screen, found at all stations of the U. S. Weather Bureau, the temperature on the side the sun is shining on will be higher than that on the other side, and since it is impossible to expose all instruments in the center of the shelter where conditions are nearly uniform, there will occur differences more or less troublesome. Differences of temperature exceeding 1.5° C. and differences of humidity exceeding 12 per cent. have been observed in extreme cases of this kind. If the instrument shelter is in an exposed place in a windy region these differences are small.

In comparing thermographs, hygrometers and hygrographs with standard instruments, differences of sensitiveness are important, and are sometimes erroneously applied as corrections. For example, in Figure 20, if the actual changes of temperature by a sensitive mercurial thermometer are represented by the irregular line a, the tracing of the more sluggish thermograph would very likely correspond to the

mean change indicated by the line b. If the standard thermometer is read at *d* or *e* or *f* the thermograph apparently will have an error of +0.8°, —1°.4 or ±0.0° respectively, the amount depending entirely upon the difference of sensitiveness. Probably the only safe way to allow for such differences is to use for a standard temperature the mean of several readings at intervals of a few seconds. Any important displacement of a record-sheet will be indicated by an excess of corrections having the same sign at the beginning or end of the record.

Anemometers.—The accuracy of records of wind-velocity depends upon several considerations: whether mean or average velocities or individual gusts are to be measured, in which case the sensitiveness of the instrument and the method of registration are important; the friction of moving parts, which is most important at low velocities; and in the instance of Robinson anemometers, upon the dimensions and proportions of the cups and supporting arms. Since the beginning, nearly all Robinson anemometers have been rated on the supposition that the cups move with one third of the wind's velocity (that is, in moving a cup through a distance of one meter the wind would travel 3 meters); but recent studies show that the factor 3.00 holds only for very small instruments making about one turn for each meter of wind, and decreases to 2.45 and 2.20, respectively, in the Weather Bureau pattern, making one rotation for each 3 meters, and the Kew pattern, of which one rotation is equal to 11.5 meters.[19] Furthermore, the factors of instruments whose cups are small relative to the length of arm are nearly constant, while those of instruments having large cups on comparatively short arms vary according to the velocity. Records made by Weather Bureau anemometers should be corrected by means of the table prepared by Professor Marvin. Corrections for Robinson instruments of other dimensions can be determined by direct comparisons with standards, or if a standard is not available, approximate factors may be based upon those given above for three different sizes of this instrument.

Nearly all anemometers properly rated can be used for the measurement of the average or mean velocity of the wind, but when it is desired to measure rapid variations or extreme velocities there should be employed only the lightest or most sensitive (usually the smallest) rotation instruments or those of other forms in which the weight of the moving parts is small compared with the surface exposed to

[19] American Meteorological Journal, April and July, 1889, February, 1891; Quarterly Journal, Royal Meteorological Society, July, 1892; Annals, Harvard College Observatory, XL, 4.

the wind.[19] In the instance of the Dines anemometer, the sensitiveness varies with the length of tubes connecting the head and the recorder; and if this length must exceed 10 meters the standard pattern must be replaced by one of which both pressure head and transmission tubing are larger.

Aerological Apparatus.—Instruments carried by kites or balloons must be very sensitive in order to record changes of condition while moving through the air; also since at great heights it is impossible to secure comparisons with standard instruments, the ranges or scale-values of the various elements need to be determined with great care under artificial conditions approximating those of the upper air. The effect, on the records of temperature and humidity, of the direct rays of the sun, which are very intense at great elevations, must be eliminated by insulating the thermometric and hygrometric elements from parts of the instrument exposed to the sun; and the anemometers should be exposed in such a manner that they are not shielded by any part of the kite or balloon and are not influenced by changes of level.[20] Under favorable conditions the records obtained from instruments of this kind are as accurate as those obtained from fixed instruments at the ground.

[19] American Meteorological Journal, April and July, 1889, February, 1891; Quarterly Journal, Royal Meteorological Society, July, 1892; Annals, Harvard College Observatory, XL, 4.

[20] Annals, Harvard College Observatory, XLII, 1; XLIII, 3.

CHAPTER III.

ATMOSPHERIC TEMPERATURE.

PART I. VERTICAL DISTRIBUTION OF TEMPERATURE.

GENERAL STATEMENT.

Long before systematic observations of free air conditions were made it was in a general way well known that, on the average, temperature decreases with altitude. During the past 50 years such observations, with increasing refinement of methods and apparatus, have not only conclusively demonstrated that this belief was well founded, but have also determined with considerable accuracy the amount of this temperature decrease to heights of 5 kilometers and with fair accuracy up to about 30 kilometers. These observations include those made at mountain stations, and in the free air by means of kites and balloons. The former have been obtained from eye readings and from self-recording instruments; the latter almost entirely from self-recording instruments, known as meteorographs (see Chapter II). Observations at mountain stations do not give accurately the conditions of the free air, because of the influence of the mountain itself through its absorption and emission of radiant energy; also in its effect upon air circulation and the courses taken by passing areas of high and low pressure. For example, the diurnal range of temperature at a mountain station, although much smaller than at a low land station, shows nevertheless a maximum in the afternoon and a minimum in the early morning (Figure 21), whereas in the free air, the times of these extremes are entirely different from those of the earth's surface (p. 42). In the annual range no such decided difference appears to exist, although even in this case the times of maximum and minimum temperatures on mountains, viz.: July and January, respectively, agree closely with those at low land stations, whereas in the free air there is considerable retardation, amounting to about one month at altitudes of 4 to 5 kilometers.

CAUSE OF TEMPERATURE DECREASE WITH ALTITUDE.

The change of temperature with altitude is primarily due to dynamic heating and cooling. By this is meant that, if air is compressed, work is done on it and its temperature is raised, and if expanded it does work and is cooled. A familiar example of heating due to compression

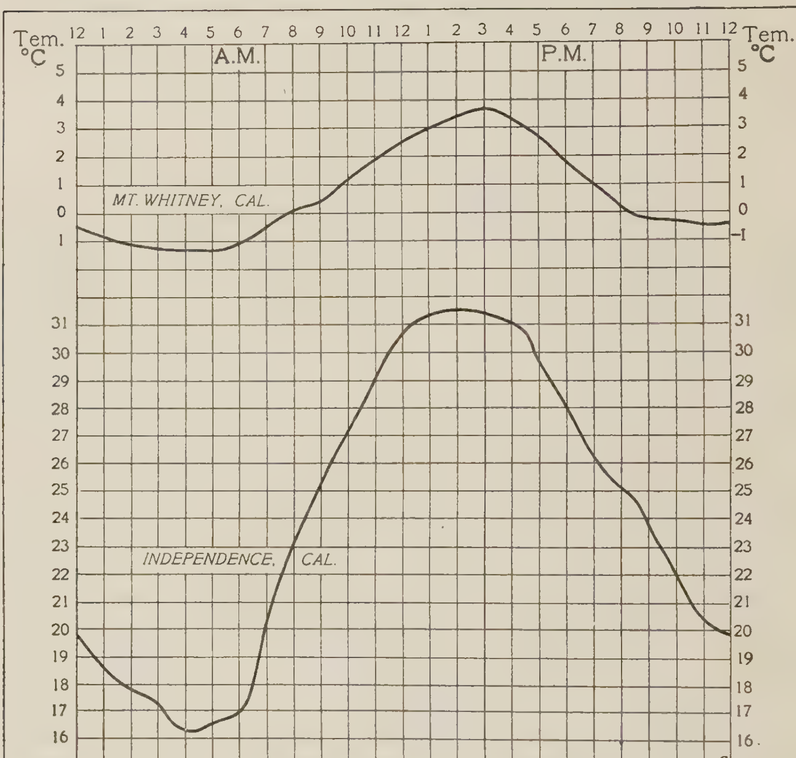

FIGURE 21. Mean diurnal temperature range at Mount Whitney (altitude 4410 m.) and Independence, Ca. (altitude 1191 m.) during the period, August 3 to 12, 1913, inclusive.

occurs when air is forced into a tire by a bicycle pump. During this process the barrel of the pump is likely to become too hot to touch. Conversely, if air expands, it gives up heat, or, in other words, becomes cool itself. As will be explained in Chapter IV, air pressure and therefore air density, diminishes with increasing altitude. Hence ascending air expands and descending air is compressed. In the case

of unsaturated air, if not materially affected by gain of heat from the surrounding air or by loss of heat to it, the expansion when rising, and contraction, when falling, are such that a change of temperature of approximately 1° C. per 100 meters change in altitude is produced. This change is known as the "adiabatic rate," and may be briefly defined as the change in temperature brought about by a change in density, or, in other words, the actual temperature of the mass of air under consideration is changed, but its potential temperature (temperature due to position) remains unchanged. Under such ideal conditions any mass of air that may be moved up or down will remain

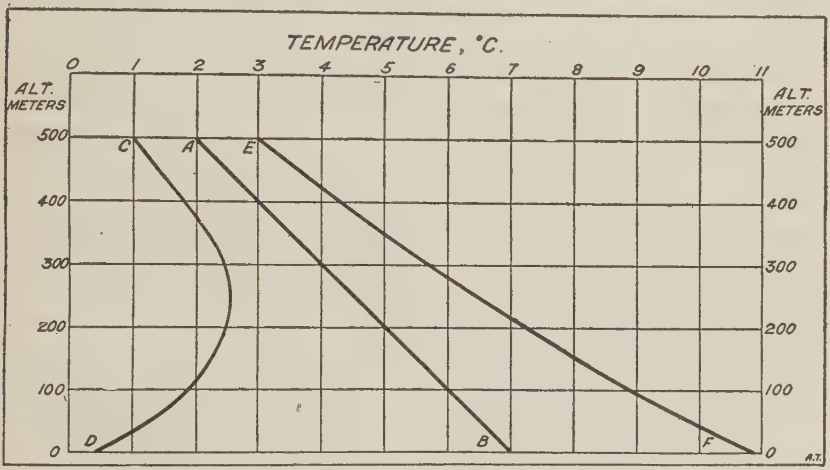

FIGURE 22. Examples of different states of equilibrium: AB, adiabatic gradient for dry air (neutral equilibrium); CD, temperature inversion (stable equilibrium); and EF, superadiabatic gradient (unstable equilibrium).

in its new position, because its condition of equilibrium, i. e., its potential energy, has not been changed. If, however, the temperature change with height, or "gradient" as it is commonly called, is less than 1° C. per 100 meters, air that is moved upward will cool to a lower temperature than that of the air with which it comes in contact and will therefore fall back to its initial level; if the gradient is greater than 1° C. per 100 meters (a condition which may occur for short distances only near the earth's surface during the daytime) the lower portions of the air will ascend and the upper portions descend until the gradient of the whole mass has returned to the adiabatic, or 1° C. per 100 meters. In these three states air is respectively in neutral, stable and unstable equilibrium, the degree of stability or

instability being proportional to the extent of the variation from the adiabatic rate. For example, during calm, clear nights the temperature is often lower at the earth's surface than above it. This condition is called an "inversion" of temperature, or, a change with altitude the opposite of that which normally occurs, viz.: decreasing temperature with increasing altitude, and is, of course, characterized by great stability. In the afternoon, on the other hand, excessive heating often produces what is called a "super-adiabatic" condition, or, a change with altitude in excess of 1° C. per 100 meters, and is marked by pronounced instability. The curves in Figure 22 illustrate these different states. AB represents the adiabatic state, or neutral equilibrium; CD, an inversion, or stable equilibrium; and EF, a super-adiabatic state, or unstable equilibrium. Examples of stable and unstable equilibrium may also be seen in Table 2.

Effects of Moisture.—As a matter of fact, the atmosphere is never entirely free from moisture and, as the specific heat of water vapor is nearly twice that of dry air, it is evident that the adiabatic rate of temperature will diminish in proportion to the amount of water vapor present; moreover, as the capacity of air[1] for water vapor is a function of its temperature, the adiabatic rate of moist air also diminishes with increasing temperature. This decrease is very small for all conditions of humidity, until saturation, i. e., condensation of the water vapor, occurs. When this state is reached, a marked decrease in the adiabatic rate takes place, due to the "latent heat of vaporization" or "latent heat of fusion" (or both, depending upon whether the temperature is above or below freezing) that is set free. This change in the adiabatic rate is well shown in Figure 23, constructed by Neuhoff. The straight diagonals represent the adiabatic rate of change for dry air, 1° C. per 100 meters; the broken lines, moisture content in grams of water vapor per kilogram of dry air; and the dot and dash lines, saturation adiabats, at the various temperatures from —30° C. to +30° C., and for altitudes from sea level to 7 kilometers. In order to grasp the significance of this effect of moisture on the adiabatic rate it is essential to know the meaning of the expression "latent heat." Briefly, it represents the energy, or work, required to change any substance from the solid to the liquid state, or from the

[1] Although it is customary to speak of the capacity of air for water vapor, in reality the air itself has no effect in this respect, except in so far as its temperature affects that of the water vapor. In other words, a cubic meter of space without air can contain exactly the same amount of water vapor as it can, if air is present, providing the temperature is the same in both cases.

liquid to the gaseous state. This change from one state to another is not accompanied by a change in temperature, the heat used being merely a form of energy by means of which the change is brought about. But, when the reverse process occurs, this stored up energy is released as heat and there is consequently a rise in temperature.

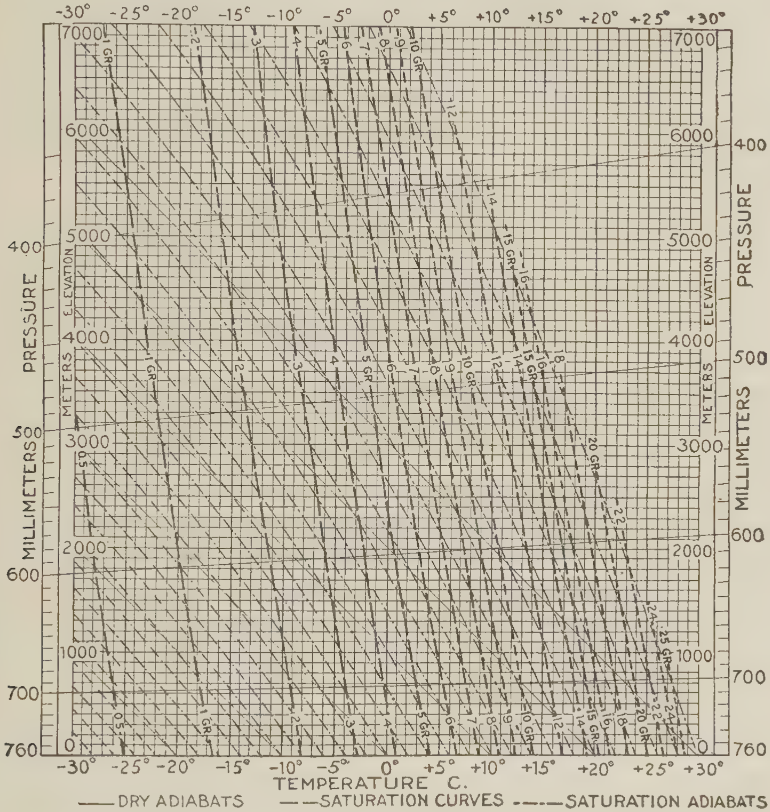

FIGURE 23. Adiabatic diagram (Neuhoff).

Thus, when water is converted into water vapor, no temperature change occurs, but, when this water vapor condenses as fog, cloud, rain, etc., the so-called latent heat is given up, and the temperature of the air rises. The energy given up when water changes from the gaseous to the liquid state is called latent heat of vaporization; when from the liquid to the solid state, latent heat of fusion; when from the gaseous directly to the solid, i. e., at temperatures below freezing, both forms are liberated.

Now air, in rising, sooner or later reaches a height at which its reduced temperature causes condensation of water vapor and consequent setting free of latent heat. Conversely, when air in which condensation has already taken place descends, it becomes warmed dynamically to a temperature at which evaporation and therefore absorption of heat occurs. In the case of descending air in which condensation does not exist, however, its temperature continues to rise until it reaches a level of the same temperature as that which it itself has acquired. Since descending air on the average is drier than ascending, because of the rain, etc., that has fallen out of the latter, the adiabatic gradient for dry air holds to a greater elevation as a rule in descending than in ascending air.

Other Influences.—Although moisture exercises the greatest effect upon temperature decrease with altitude, there are certain other causes which should be briefly considered. Of these the most important are the character of the earth's surface, or topography, and the movements of passing areas of high and low pressure, with their attendant changes in temperature, cloudiness, etc.

Effects of Topography.—Evaporation from land areas usually is less than from water surfaces. Moreover, the specific heat of land surfaces is relatively low. Hence they become heated to a much greater extent during the day and summer and cool to a lower temperature during the night and winter. In addition, the character of the earth's surface exerts a considerable influence, black soils absorbing and radiating more readily than light colored soils. These effects are large at and near the earth's surface, but relatively small at heights of a kilometer above it (p. 51).

Effects of Passing Highs and Lows.—The movements of Highs and Lows bring to a locality marked changes in temperature and moisture which sometimes extend to great altitudes (Chapter IX). The changes are, however, greatest at the surface for the most part and therefore the vertical temperature gradient undergoes considrable alteration, being least during extreme cooling at the earth's surface, and greatest during extreme heating. For example, a winter anticyclone, i. e., an area of high pressure, is usually accompanied by abnormally low temperatures at the earth's surface with a marked inversion or increasing temperature with increasing altitude at higher levels. On the other hand, a cyclone, i. e., an area of low pressure, is relatively warm at the surface, but the temperatures ordinarily diminish with altitude, at any rate up to the top of the cloud layer, above which there is usually a moderate inversion.

MEAN TEMPERATURE GRADIENTS IN THE LOWER ATMOSPHERE.

Annual Range.—The most systematic and complete exploration of the air up to about 5 kilometers has been made by means of kites. These observations have been obtained principally in the United States and Europe and are well distributed throughout the year. At higher

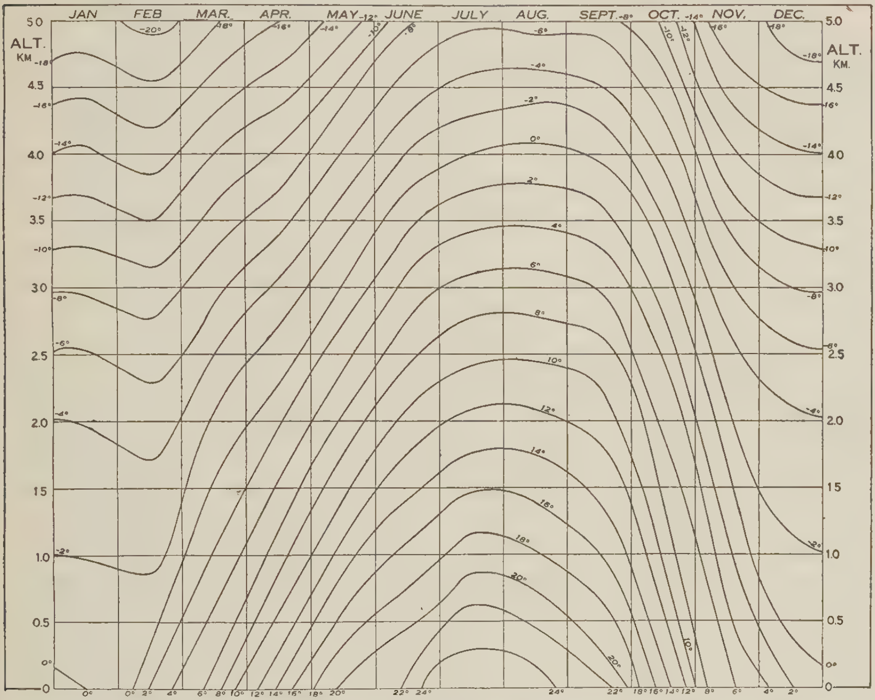

FIGURE 24. Mean annual march of free air temperatures, °C, above Mount Weather, Va., altitude 526 meters. (Lines extended to sea level by extrapolation.)

levels observations have been made almost wholly by means of sounding balloons. These are fewer in number, but, on the other hand, smaller variations occur in the temperature gradients at great heights and therefore a smaller number of observations suffices to give very satisfactory information at those altitudes. In the discussion of temperatures up to 5 kilometers, greatest weight will be given to the kite observations, because of their greater number and better distribution. Figure 24 shows the annual march of temperature as observed at

Mount Weather, Va., during the five years July 1, 1907, to June 30, 1912, inclusive.[2] An examination of this figure indicates that the annual range is greatest at the earth's surface, 526 meters above sea level, and decreases to a little more than half its surface value at the 5 kilometer level. In the figure, the lines have been extended by extrapolation to sea level[3] and, if these values are considered, it is seen that the annual range at this and the 5 kilometer level is in almost the exact proportion of 2 to 1. The reason for this is readily apparent in the relatively small gradients in the lower levels during the winter as compared with those during the summer. It should not, of course, be understood that this relation will apply at all places. For example, the annual range at Drexel, Nebr., at both the surface and the 5 kilometer level is much larger than at Mount Weather, because of the continental character of its climate, and the proportion in the range at the surface and the 5 kilometer level is somewhat greater than two to one. In Europe, on the other hand, due to the effect of the Atlantic Ocean, the annual range at both levels is comparatively small and the proportion approximately six to five. In tropical regions[4] the annual range is less than 2° C. both at sea level and at the 5 kilometer level. Another striking fact brought out by the figure is the rapid cooling at all levels during the month of October in contrast with the relatively slow warming during the spring months, due to the greater diathermance of the atmosphere, i. e., its ability to transmit heat during the former than during the latter period. The surface maximum and minimum temperatures occur respectively in July and January. At higher levels, especially above 3 kilometers, there is a retardation of about one month in these extremes.

Table 1 contains the average seasonal and annual temperature gradients, in degrees centigrade per 100 meters as observed at Mount Weather, Va.,[5] Drexel, Nebr.,[6] Blue Hill Observatory, near Boston, Mass.,[7] and at several stations in Europe.[8] The figures given show

[2] Bulletin of the Mount Weather Observatory, Vol. 6, pp. 111-194; also Monthly Weather Review, January, 1918, pp. 11-20.

[3] For values at various heights from sea level to 7 kilometers see Monthly Weather Review, January, 1918, pp. 11-20.

[4] Results of Registering Balloon Results at Batavia, by van Bemmelen.

[5] Bulletin of the Mount Weather Observatory, Vol. 6, part 4.

[6] Not yet published; based on 2½ years' observations.

[7] Annals of the Astronomical Observatory, Harvard College, Vol. LVIII, part 1, p. 59.

[8] Die Temperaturverhältnisse in der freien Atmosphäre. Wagner, in Beiträge zur Physik der freien Atmosphäre, Band 3. Heft 2-3.

TABLE I.

Mean seasonal and annual temperatures and temperature gradients at Mount Weather, Va., Drexel, Nebr., Blue Hill, Mass., and in Central Europe.

Altitude, Sea Level. Meters.	Mt. Weather, Va. Mean °C	Mt. Weather, Va. Δt/100m °C	Drexel, Nebr. Mean °C	Drexel, Nebr. Δt/100m °C	Blue Hill, Mass. Mean °C	Blue Hill, Mass. Δt/100m °C	Europe Mean °C	Europe Δt/100m °C
				SPRING.				
aSurface	....	....	10.3	....	5.6	....	5.6	....
500	10.8	....	9.6	0.67	4.3	0.43	....	....
750	9.2	0.64	7.9	0.68	....	....	....	....
1000	7.8	0.56	6.8	0.44	1.7	0.52	1.6	0.40
1250	6.5	0.52	5.8	0.40	....	....	....	....
1500	5.2	0.52	4.8	0.40	—0.8	0.50	....	....
2000	2.6	0.52	2.5	0.46	—2.7	0.38	—3.4	0.50
2500	0.0	0.52	—0.2	0.54	—4.6	0.38	....	....
3000	—2.8	0.56	—3.0	0.56	—7.1	0.50	—8.8	0.54
3500	—5.8	0.60	—5.9	0.58	—9.8	0.46	....	....
4000	—8.8	0.60	—8.8	0.58	....	....	—14.6	0.58
4500	—12.1	0.66	—12.1	0.66	....	....	....	....
5000	—15.5	0.68	—15.4	0.66	....	....	—21.4	0.68
				SUMMER.				
aSurface	....	....	23.7	....	18.9	....	15.7	....
500	21.6	....	23.0	0.67	17.5	0.46	....	....
750	19.8	0.72	21.4	0.64	....	....	....	....
1000	18.2	0.64	20.1	0.52	14.8	0.54	11.4	0.43
1250	16.6	0.64	18.5	0.64	....	....	....	....
1500	15.1	0.60	17.0	0.60	11.9	0.58	....	....
2000	12.1	0.60	13.7	0.66	9.1	0.56	6.1	0.53
2500	9.3	0.56	10.2	0.70	6.4	0.54	....	....
3000	6.4	0.58	6.8	0.68	3.7	0.54	0.9	0.52
3500	3.2	0.64	3.4	0.68	0.8	0.58	....	....
4000	—0.1	0.66	0.2	0.64	....	....	—4.6	0.55
4500	—3.4	0.66	—3.1	0.66	....	....	....	....
5000	—6.4	0.60	—6.3	0.64	....	....	—10.7	0.61
				AUTUMN.				
aSurface	....	....	11.7	....	9.2	....	10.9	....
500	12.1	....	11.3	0.38	8.0	0.39	....	....
750	10.7	0.56	10.2	0.44	....	....	....	....
1000	9.4	0.52	9.2	0.40	5.3	0.54	7.5	0.34
1250	8.3	0.44	8.4	0.32	....	....	....	....
1500	7.3	0.40	7.6	0.32	3.1	0.44	....	....
2000	5.6	0.34	5.4	0.36	1.5	0.32	2.6	0.49
2500	3.5	0.42	2.8	0.52	—0.4	0.38	....	....
3000	0.9	0.52	0.0	0.56	—3.3	0.58	—2.3	0.49
3500	—2.0	0.58	—2.8	0.56	—6.1	0.56	....	....
4000	—5.0	0.60	—5.5	0.54	....	....	—7.9	0.56
4500	—8.1	0.62	—8.0	0.50	....	....	....	....
5000	—11.2	0.62	—10.6	0.52	....	....	—13.8	0.59

TABLE I.—*Con't.*

Altitude, Sea Level. Meters.	Mt. Weather, Va. Mean °C	Δ^t/100m °C	Drexel, Nebr. Mean °C	Δ^t/100m °C	Blue Hill, Mass. Mean °C	Δ^t/100m °C	Europe Mean °C	Δ^t/100m °C
				WINTER.				
[a]Surface	....	....	—6.4	....	—4.0	....	—0.5	....
500	—0.7	....	—6.6	0.19	—5.4	0.46	....	....
750	—1.5	0.32	—6.5	—0.04	....	....	....	....
1000	—2.1	0.24	—5.6	—0.36	—7.4	0.40	—2.1	0.16
1250	—2.6	0.20	—4.7	—0.36	....	....	....	....
1500	—3.0	0.16	—4.6	—0.04	—9.4	0.40	....	....
2000	—4.2	0.24	—5.3	0.14	—10.6	0.24	—5.1	0.30
2500	—6.0	0.36	—7.1	0.36	—12.4	0.36	....	....
3000	—8.4	0.48	—9.5	0.48	—14.8	0.48	—9.6	0.45
3500	—11.2	0.56	—11.9	0.48	—17.5	0.54	....	....
4000	—13.9	0.54	—14.6	0.54	....	....	—15.5	0.59
4500	—16.8	0.58	—17.6	0.60	....	....	....	....
5000	—19.7	0.58	—20.8	0.64	....	....	—21.7	0.62
				YEAR.				
[a]Surface	....	....	9.8	....	7.4	....	7.9	....
500	11.0	....	9.3	0.48	6.1	0.43	....	....
750	9.6	0.56	8.2	0.44	....	....	....	....
1000	8.4	0.48	7.6	0.24	3.6	0.50	4.6	0.33
1250	7.3	0.44	7.0	0.24	....	....	....	....
1500	6.2	0.44	6.2	0.32	1.2	0.48	....	....
2000	4.0	0.44	4.1	0.42	—0.7	0.38	0.1	0.45
2500	1.7	0.46	1.4	0.54	—2.8	0.42	....	....
3000	—1.0	0.54	—1.5	0.58	—5.4	0.52	—5.0	0.51
3500	—4.0	0.60	—4.4	0.58	—8.2	0.56	....	....
4000	—7.0	0.60	—7.3	0.58	....	....	—10.7	0.57
4500	—10.2	0.64	—10.3	0.60	....	....	....	....
5000	—13.3	0.62	—13.3	0.60	....	....	—16.9	0.62

[a] Mount Weather, 526 meters above sea level. Drexel, 396 meters above sea level. Blue Hill, 195 meters above sea level. Europe, observations reduced to sea level.

that mean gradients agree fairly closely in different parts of the world, except near the surface, where local influences are most pronounced. For example, in the winter season, there is on the average an inversion layer, i. e., one in which the temperature increases with altitude, at Drexel extending to the 1500 meter level. This is due to the fact that this station has a typical continental climate and is, moreover, in the path of the greater number of strong anticyclones, which as before stated are characterized by extreme cold at the surface and in general by an inversion above it. At Mount Weather, Blue Hill and in Europe the winter gradient is positive, but very small. In

the other seasons the gradients are much steeper at all places. For the year the mean value from the surface to the 5 kilometer level is approximately 0.5° C. per 100 meters, or about half the adiabatic

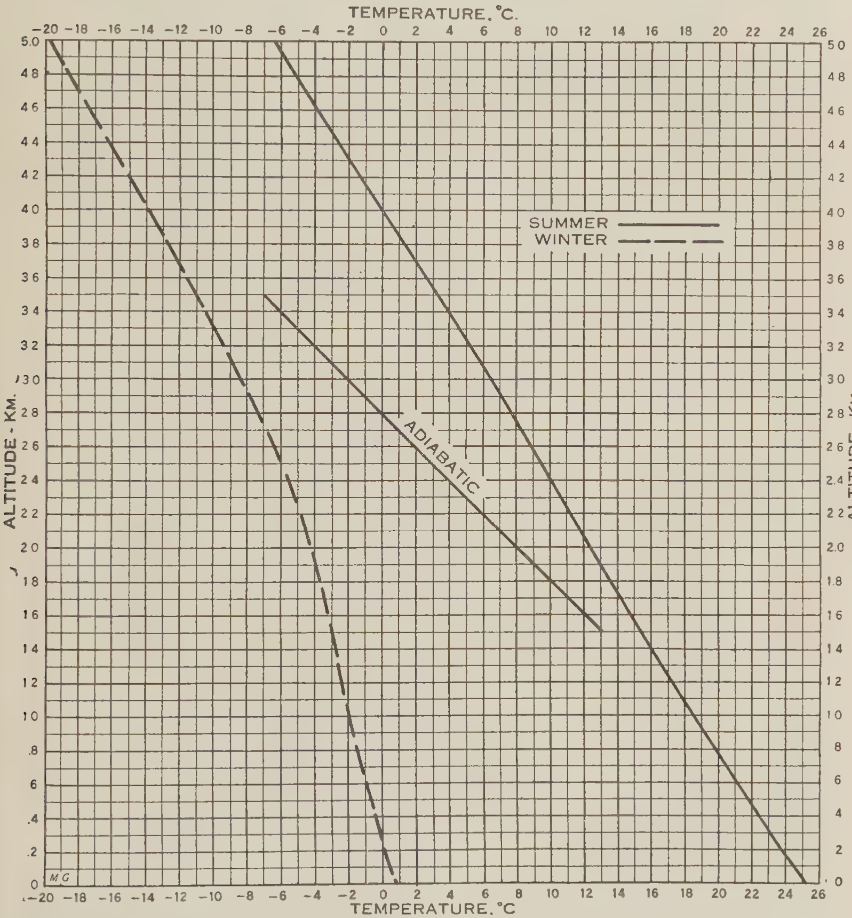

FIGURE 25. Mean summer and winter free air temperature gradients, °C, above Mount Weather, Va. (Extended to sea level.)

rate for dry air. This value increases somewhat with altitude, being practically 0.6° C. per 100 meters between 3 and 5 kilometers. The mean summer and winter gradients, as observed at Mount Weather, are shown in Figure 25. Values between the surface and sea level have been estimated by extrapolation.

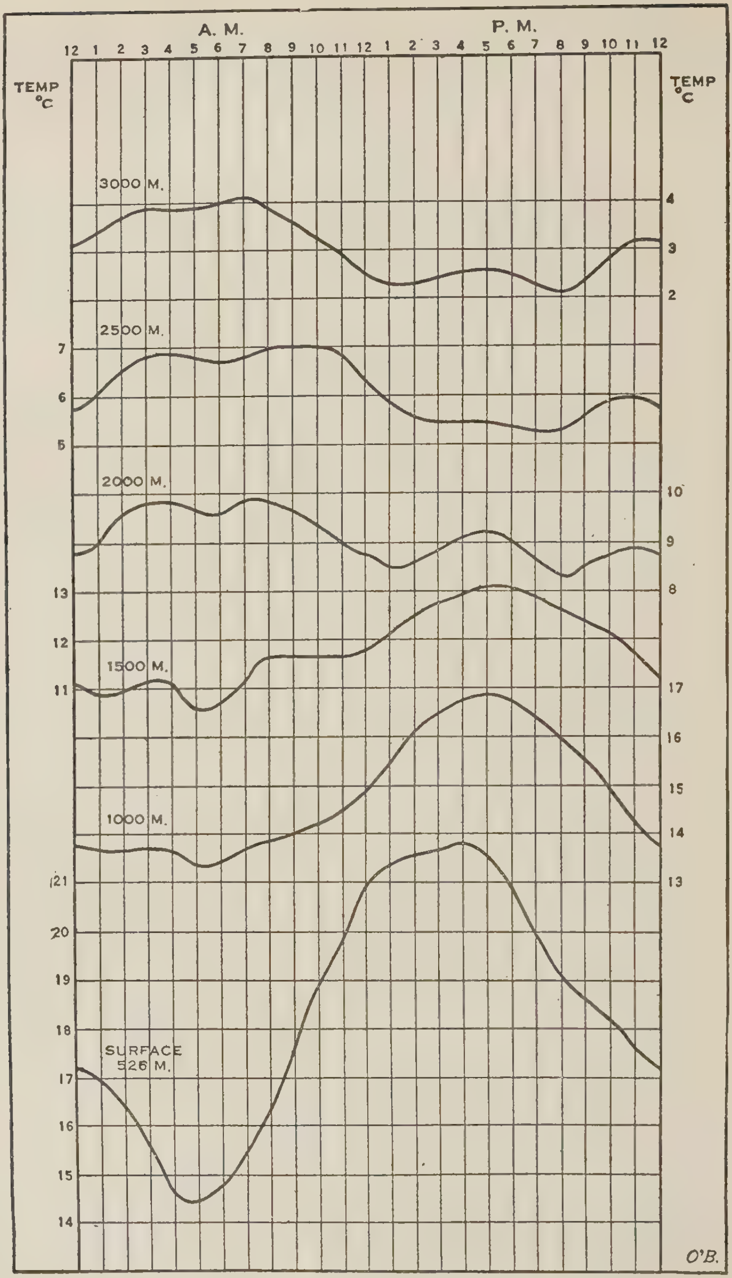

FIGURE 26. Diurnal distribution of temperature for the summer half of the year at different levels above Mount Weather, Va.

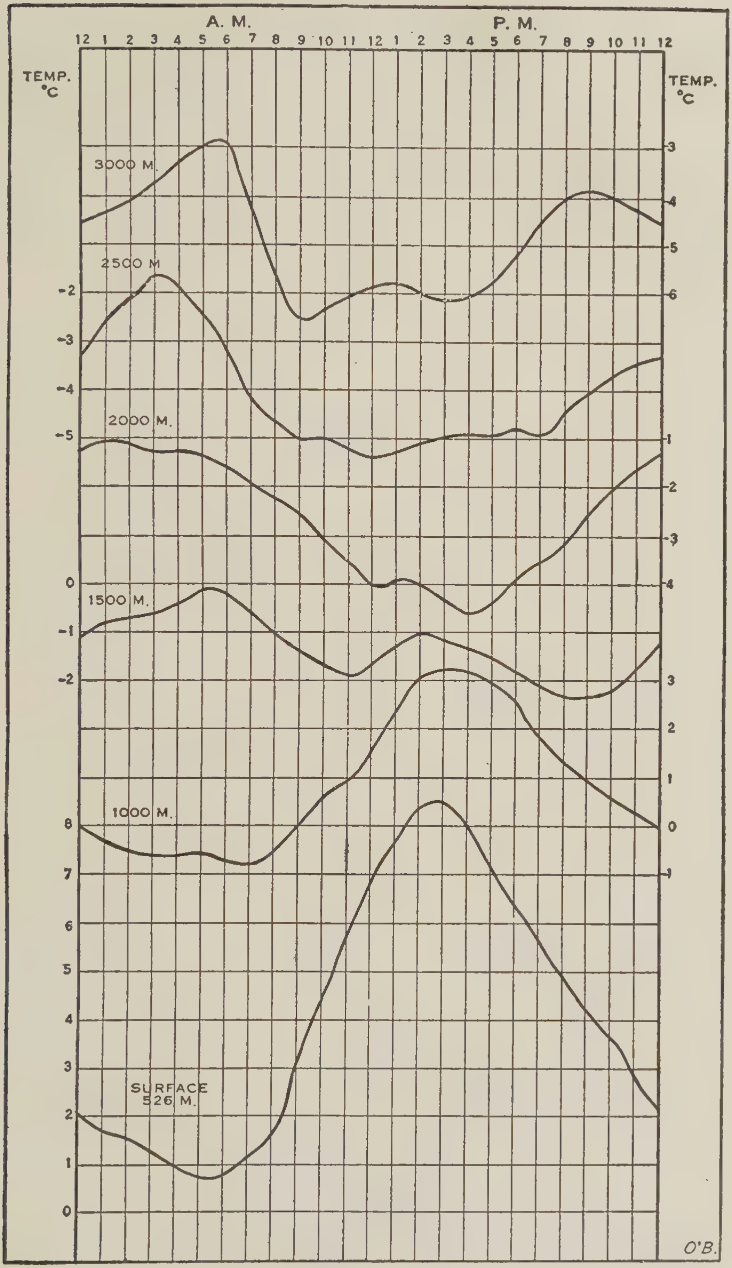

FIGURE 27. Diurnal distribution of temperature for the winter half of the year at different levels above Mount Weather, Va.

Diurnal Range.—Observations of diurnal range of temperature and other elements at different altitudes have been made at Mount Weather, Va.,[9] and at Drexel, Nebr. These observations cannot be successfully made except under certain favorable conditions of wind and weather. They may be said to represent fairly well conditions during relatively clear weather, when radiation is most active and irregularities due to changing winds, storms, etc., are least in evidence. Figures 26 and 27 show the average diurnal range of temperatures for the summer and winter halves of the year at certain specified levels, as observed at Mount Weather. The lowest curve in each figure shows the characteristic diurnal range at the surface, viz.: maximum temperature about 3 to 4 P. M. and minimum temperature about 4 to 6 A. M., the exact time of these extremes depending on the season. At the 1000 and 1500 meter levels in summer and at the 1000 meter level in winter this relation still persists, although the range is much smaller. The surface effect is felt to a greater height in summer than in winter because of the greater convectional activity during that season. Above these levels a reversal occurs: the maximum temperatures occur in the early morning and the minimum temperatures during the afternoon. The explanation of this is fairly simple. During the daytime the earth's surface is heated by solar radiation and the air in contact with it is forced to rise. As it rises, more or less condensation occurs in the form of detached cumulus clouds and the water vapor in these clouds absorbs whatever terrestrial radiation there may be with the result that but little of this radiation reaches the higher levels. Air being a poor absorber of solar radiation, little heat is gained from that source. During the night, on the other hand, rapid cooling occurs at the earth's surface, the air in the cloud layer descends somewhat and in so doing is heated dynamically with the result that evaporation takes place. Thus a certain amount of terrestrial radiation is able to pass through the layer in which during the day there was condensed water vapor and this radiation is to a certain extent absorbed by the air at higher levels and produces the nocturnal maximum. Observations at altitudes above 3 kilometers are not sufficiently numerous to determine the height to which this diurnal effect extends, but in one series at Fort Omaha, Nebr., with sounding balloons[10] there is evidence to show that it is not felt above 5 or 6 kilometers. Referring again to figures 26 and 27, it is apparent

[9] Bulletin of the Mount Weather Observatory, Vol. 6, part 5.
[10] Monthly Weather Review, Vol. 44, pp. 247-264.

that between the surface and the 1000 meter level there is a very small vertical gradient during the night and a very large one during the afternoon. If we consider these values in greater detail, i. e., by shorter altitude intervals, we find that the gradients amount to a strong inversion at about 100 meters above the surface at night and a super-adiabatic condition in the afternoon, the latter extending to about 400 meters. Table 2 shows mean gradients for summer and winter as observed at 2 A. M. and 2 P. M. at and near Mount Weather, Va., during clear weather.

TABLE II.

Mean Temperature Gradients during clear weather observed at and near Mount Weather, Va.

Altitude above Surface. Meters.	Summer 2 a. m. $\Delta^t/100m.$	Summer 2 p. m. $\Delta^t/100m.$	Winter 2 a. m. $\Delta^t/100m.$	Winter 2 p. m. $\Delta^t/100m.$
0	....	....	....	....
100	—3.40	1.60	—6.90	1.30
200	0.20	1.20	0.60	1.10
300	0.60	1.10	0.60	1.10
400	0.60	1.10	0.60	1.00
500	0.70	0.90	0.60	1.00
750	0.60	0.88	0.48	0.88
1000	0.56	0.80	0.24	0.96

From this table it is easy to see that conditions at night are characterized by marked stability, whereas in the daytime the air is in a state of unstable equilibrium, resulting in strong convectional activity. These conditions are most pronounced during clear, quiet weather, when the relative humidity is low. When the humidity is high, especially during the summer half of the year, thunderstorms are likely to result.

Observations at Heights above 5 Kilometers.—As already stated, extensive exploration of the air to great heights has been conducted by means of sounding balloons, which carry small self-recording instruments. These observations have been made for the most part in the United States, Europe, Canada, Australia and Java (Batavia). The most important result of these investigations is the discovery of a region in the upper atmosphere in which the temperature ceases to diminish with increase of altitude, and in fact has a tendency to increase to some extent. This layer has been variously called the "Isothermal Region," "Advective Region" and "Stratosphere," but is now generally known by the last given name. Below its base and

extending down to the 5 kilometer level above the earth's surface, the temperature gradient is approximately 7° C. per kilometer and does not vary greatly for different places or seasons.[11] The height of the base of the stratosphere does vary, however, with latitude, season and surface pressure conditions. The seasonal variation may, in a sense, be considered a latitude effect, the height diminishing in proportion to the distance of any given locality from the thermal equator. In Table 3 may be found the average summer, winter and annual heights of the base of this region as observed at different places.

Table III.

Average Summer, Winter and Annual Altitudes of the Base of the Stratosphere.

Place	Latitude	Altitude Summer Km.	Altitude Winter Km.	Altitude Annual Km.
Omaha, Nebr.	41	12.5	10.0	11.2
St. Louis, Mo.	39	12.0	10.5	11.2
Avalon, Cal.	33	16.0	....	....
Southern Canada	45	13.0	10.0	11.5
Southern England	51	....	....	10.5
Continental Europe	50	....	....	11.0
Melbourne, Australia	38	....	....	10.0
Batavia, Java	7	....	....	17.0

Associated with this difference in height of the stratosphere is a corresponding difference in its temperatures, these being lowest as a rule when the altitude of the base of the stratosphere is greatest. Thus, in the equatorial regions they are somewhat lower than —80° C. and do not vary materially throughout the year. As greater distances from the equator are reached, the stratosphere is found to have a lower altitude and a higher temperature. In middle latitudes, the annual range in the height is, roughly, 2.5 kilometers, but in this case observations do not agree as to the relation of temperature to this seasonal change in height. Thus, European observations give, on the average, a higher temperature in summer than in winter, whereas the reverse of this has been found to exist in Canada. In the United States,[12]

[11] As stated earlier in this chapter the mean gradient from the earth to the 5 kilometer level is about 5° C. per kilometer. This gives a mean gradient, from the earth's surface to the base of the stratosphere, of 6° C. per kilometer approximately.

[12] For seasonal values at various altitudes from sea level to 32 kilometers, see Monthly Weather Review, January, 1918, pp. 11 to 20.

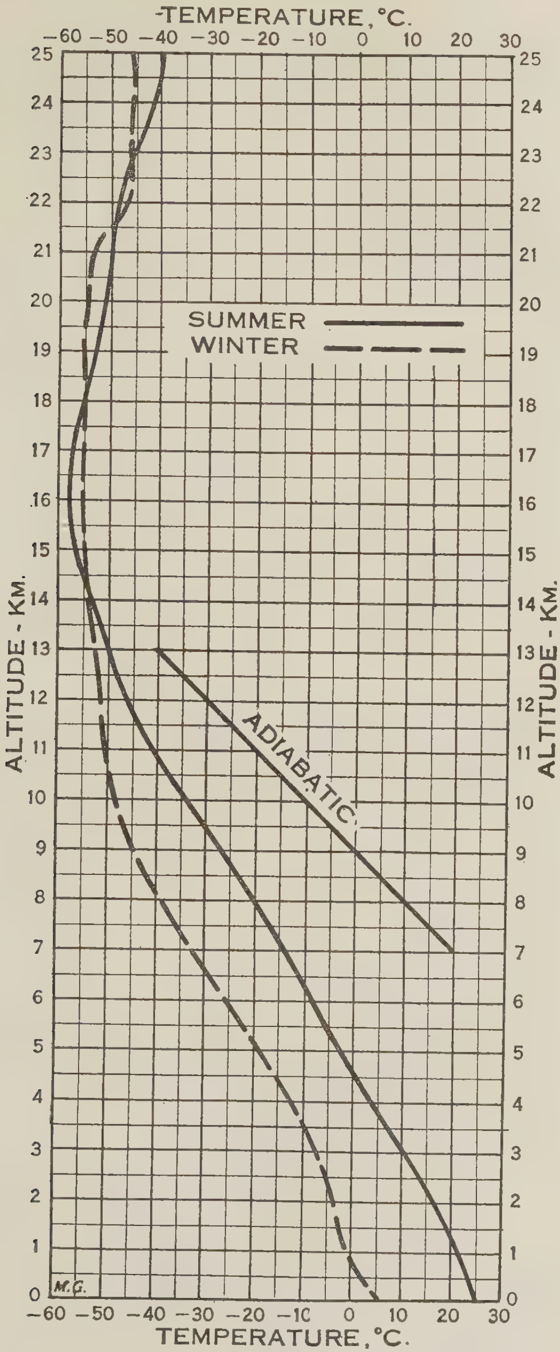

FIGURE 28. Mean summer and winter free air temperature gradients, °C, from sounding balloon records in the Central and Western States.

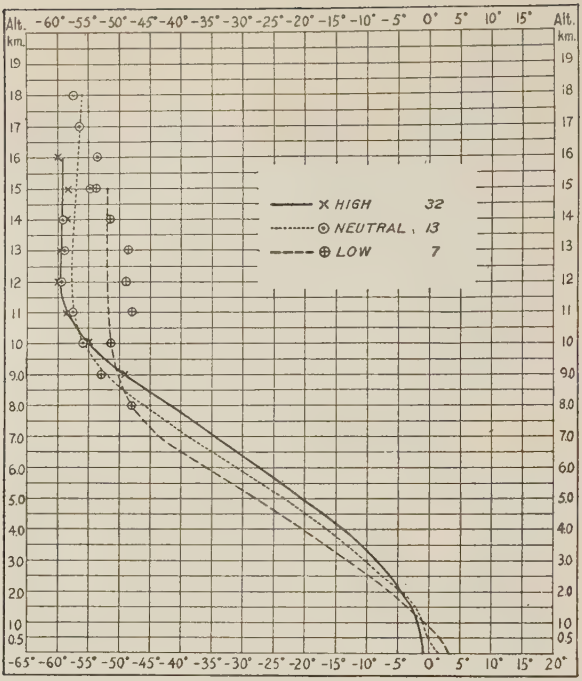

FIGURE 29. Temperature gradients, °C, at different pressures in winter, as observed in Europe.

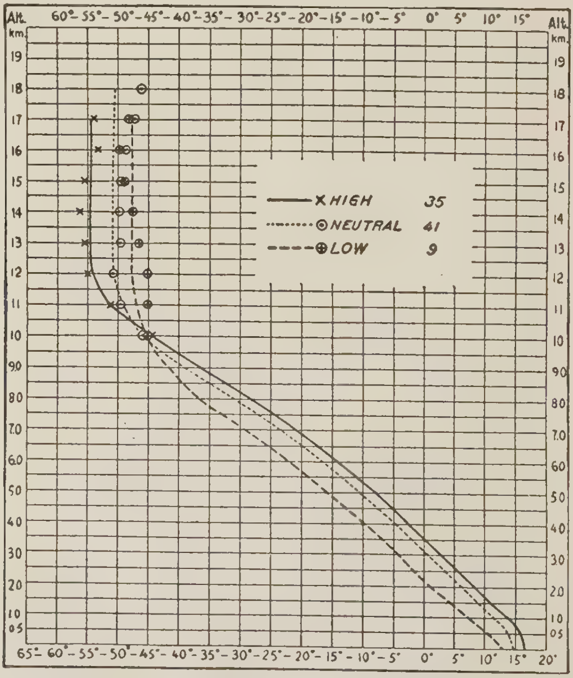

FIGURE 30. Temperature gradients, °C, at different pressures in summer, as observed in Europe.

as shown in Figure 28, there is little difference in the mean temperatures of the two seasons. The reason for these variations is not clear and more observations are needed to establish them definitely. If

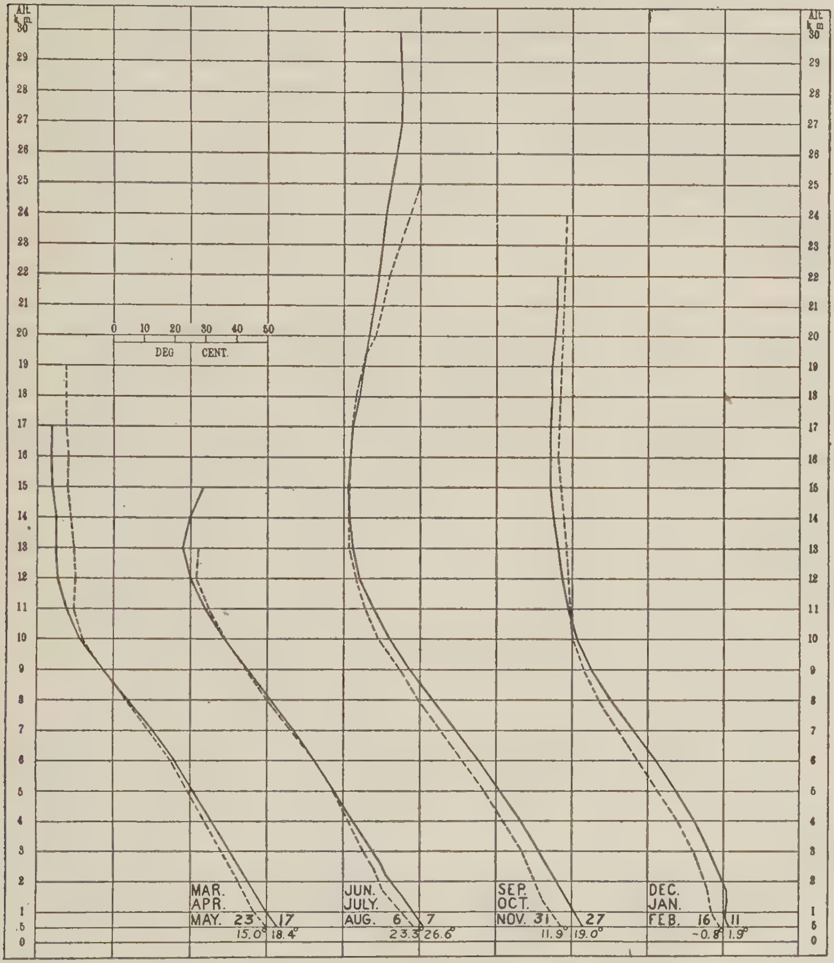

FIGURE 31. Temperature gradients, °C, over falling (solid lines) and over rising (broken lines) air pressure at the earth's surface, as observed in the United States.

they really do exist, it may be that the greater frequency of high cirrus clouds in some regions than in others and at one season than at another is the cause. Such clouds would interfere with the transmission of terrestrial radiation to the stratosphere, a case somewhat analogous

to the effect of lower clouds upon the diurnal range at different levels, as pointed out on page 42.

Vertical Temperature Gradients Related to Surface Pressure.—In "Physics of the Air," Chapter IV, Humphreys has grouped a large number of observations, obtained at four stations in Europe, according to the surface pressure conditions prevailing at the times of the ascensions. The results are indicated in Figures 29 and 30 and are, briefly, as follows: (1) the altitude of the base of the stratosphere is greater when pressure at the earth's surface is high than when it is low; (2) except at and near the earth's surface in winter, and even including these levels in summer, the troposphere[13] is warmer and the stratosphere colder during high than during low barometric pressure at the earth's surface. The second statement is true particularly for the central portions of high and low pressure areas, but does not apply equally well to the regions more distant. In fact, the opposite is true in the eastern and southern quadrants of highs and lows, these reversals increasing with the distance from their centers, due to the wind circulation about them. This relation of temperature to increasing and decreasing pressure at the earth's surface is well shown in Figure 31, based on sounding balloon observations made in the United States.[14] The solid lines represent mean temperatures over falling pressures, i. e., between a receding high and an approaching low; and the dashed lines, over rising pressure, i. e., between a receding low and an approaching high. Table 4 gives similar results, as determined by means of kites to an altitude of 4 kilometers.

TABLE IV.

Mean seasonal free air temperature over falling and rising air pressure, as observed at Mount Weather, Va.

	Spring		Summer		Autumn		Winter	
Altitude Meters.	Falling Pressure.	Rising Pressure.	Falling Pressure.	Rising Pressure.	Falling Pressure.	Rising Pressure.	Falling Pressure.	Rising Pressure.
	°C	°C	°C	°C	°C	°C	°C	°C
526	10.2	8.2	20.9	19.8	13.2	9.3	—0.3	1.0
1000	8.4	4.5	17.8	16.6	11.5	5.7	0.0	—3.8
2000	4.0	—0.5	12.0	11.1	7.4	2.0	—1.6	—6.2
3000	—1.6	—5.4	6.6	5.6	2.8	—2.5	—6.4	—10.6
4000	—7.6	—10.8	0.5	0.2	—3.5	—8.2	—12.0	—15.6

[13] "Troposphere" is the name given to the lower part of the atmosphere, in which, on the average, temperature decreases with altitude, as distinguished from "stratosphere," the upper part of the atmosphere, which is characterized by nearly isothermal conditions.

[14] Bulletin of the Mount Weather Observatory, Vol. 4, part 4.

The explanation of the higher temperatures, in the troposphere, over falling than over rising pressure is found in the southerly component in the winds under the former conditions and the northerly component under the latter. These components are found to persist to great altitudes, although less decidedly at higher than at lower levels.

PART II. HORIZONTAL DISTRIBUTION OF TEMPERATURE.

GENERAL STATEMENT.

The earth receives practically all of its heat from the sun, and since the atmosphere is heated to some extent by incoming solar radiation and to a still greater extent by outgoing terrestrial radiation, it follows that insolation (the name generally given to the sun's radiant energy or heat) is entirely responsible for the heat of the earth's atmosphere. It is true that some heat is received from the interior of the earth and from other heavenly bodies than the sun, but the amount of this heat is too small to be considered. It has been estimated to be insufficient to change atmospheric temperatures by more than 0.2° C.

AMOUNT OF INSOLATION.

The immense amount of radiant energy or heat emitted by the sun can be realized when we consider that the earth receives only about one two-billionth part of it. All the rest, except similar small proportions that are intercepted by the other planets, is lost in space. In spite of this constant giving up of energy, no diminution has been observed during the time that measurements have been made, and there is every reason to believe that the heat received by the earth will be practically constant for ages to come. In the course of millions of years, however, this energy will gradually diminish and eventually will cease altogether.

Variation with Distance from Sun.—The amount of insolation received by the earth as a whole varies slightly during the year because the earth's orbit, or path around the sun, is not a circle but is slightly elliptical. Its eccentricity is small, so that, the sun being at one focus, the earth is only about three million miles nearer the sun at one time

than another. The average distance is nearly ninety-three million miles. The time of nearest approach, or perihelion, occurs near the beginning of the calendar year (January 2 in 1919), and the time of greatest distance, or aphelion, near the middle of the year (July 2 in 1919). The result is that the earth as a whole is receiving most heat during the winter and least during the summer of the northern hemisphere. The difference amounts to about 7 per cent., or, in other words, is inversely proportional to the square of the distance from the sun.

Seasonal Variation with Latitude.—Because of the inclination of the earth's axis to the plane of the ecliptic and, further, because this axis always remains parallel to itself as the earth revolves around the sun, it follows that the various parts of the earth's surface receive the sun's rays at different angles in the course of each year. This inclination of the axis amounts to 66½°, or, in other words, it is 23½° from the vertical (the vertical being regarded as a line perpendicular to the plane of the ecliptic). The change thus brought about in the presentation of the earth to the sun causes an apparent migration of the latter through 47° and gives rise to our seasons. Thus, the sun's rays fall perpendicularly at the tropic of Cancer on June 21, at the equator on September 23 and March 21 and at the tropic of Capricorn on December 21. This migration gives rise to three important results: (1) The thermal equator, as distinguished from the geographical equator, moves northward and southward with the sun and materially affects the general planetary circulation (see Chapter VIII); (2) The sun's rays fall more and more obliquely on the surface of the northern hemisphere, as the sun moves southward, and vice versa in the southern hemisphere, thus varying the amount of energy per unit of surface, not only because these rays are spread out more, as their angle of incidence diminishes, but also because somewhat greater absorption occurs by reason of their longer path through the atmosphere; and (3) the relative lengths of day and night change greatly throughout the year. The amount of this change is shown in the following table:

Latitude	0°	17°	41°	49°	63°	66½°	67°21′	69°51′	78°11′	90°
Duration	12 hrs.	13 hrs.	15 hrs.	16 hrs.	20 hrs.	24 hrs.	1 mo.	2 mo.	4 mo.	6 mo.

A diagrammatic representation of the relation of insolation to season and latitude is given in Figure 32, reproduced from Davis' Elementary Meteorology. With these general considerations in mind we can now take up the various ways in which insolation, or the sun's

radiant energy, affects the earth's atmosphere. The most important of these are: transmission and absorption of solar and terrestrial radiation; conduction and convection.

Transmission and Absorption.—Only part of the insolation intercepted by the earth becomes effective in heating the earth's surface or its atmosphere. Observations by Abbot and Fowle[15] show that about 37 per cent. of it is reflected by clouds, by the surface of the earth and to a slight extent by the atmosphere itself. A large proportion of the remaining 63 per cent, is transmitted by the atmosphere

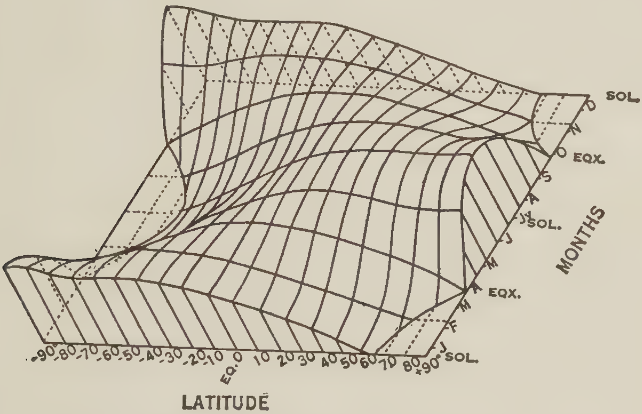

FIGURE 32. Relation of insolation to season and latitude.

directly to the earth's surface without loss by absorption. What little absorption does take place is almost altogether due to the water vapor present, although a small amount may be taken up by carbon dioxide, ozone and dust particles. There is practically no absorption by pure, dry air. Much of the insolation that reaches the earth's surface is absorbed and then reradiated. This so-called terrestrial radiation of long wave-lengths, or "heat rays," is more readily absorbed by the atmosphere than is the solar radiation of short wave lengths, or "light rays," and, in fact, is responsible in large measure for the temperature of the air. The extent of this absorption depends upon the amount of water vapor present and therefore varies greatly from time to time in its effect at various levels.

Conduction and Convection.—Air in contact with the earth's surface is heated or cooled by conduction, and therefore its temperature varies in the same sense as that of the earth's surface, although not neces-

[15] Annals Astrophysical Observatory, Smithsonian Institution, 2, p. 163.

sarily to the same extent. Conduction from one layer of air to another also occurs, but the amount of this is slight. As soon as the air next the earth's surface becomes heated, its density diminishes and it is forced by the denser portions of air above or adjacent to it to rise until it reaches a level where its charged density, i. e., the density it acquires as it rises, is the same as that of the surrounding air. This process is called convection and is one of the most important agents in the vertical distribution of temperature. It may equally well be said to be one of the most important, if not the most important, in determining the horizontal distribution of temperature over the earth.

Land and Water Surfaces in Relation to Reflection, Transmission and Absorption.—Water surfaces reflect about 40 per cent. of the insolation that reaches them, and transmit the rest to lower depths and where it is absorbed. Much of the heat that is absorbed, however, is used in evaporating the water and is therefore stored up as latent heat; some of the remainder is distributed by the constant movement of the water. The net result is that the water surface, and therefore the air above it, maintains a relatively constant temperature. Land areas, on the other hand, reflect and transmit very little insolation and there is practically no evaporation. The specific heat of land is low, and, moreover, there is no movement, as in the case of water, whereby the heat received can be convectionally distributed either horizontally or vertically. Hence, land areas become strongly heated during insolation and similarly cooled in its absence. The nature of the soil and the vegetation on it have considerable effect in this connection. A ploughed field becomes hotter than a green pasture, and shale undergoes greater temperature changes than does limestone.

Distribution of Temperature over the Earth.—The combined results of the various influences thus far discussed can best be seen if we examine a few typical charts that give mean values based upon widespread and, in some cases, long continued observations. Figure 33 shows average annual temperatures over a large part of the earth's surface. The effect of the obliquity of the sun's rays outside of the tropics is apparent in the decrease of temperature toward the poles; the effect of land areas, in the curving of the isotherms poleward over continents and in the location of the thermal equator. Charts for the summer and winter seasons, not reproduced here, show the relative influences of land and water to even better advantage. In the northern, or land, hemisphere, the isotherms are crowded closely together in winter, and spread apart widely in the summer, whereas, in the southern, or water, hemisphere, their annual variation is small. In the tropics this variation is, of course, least of all.

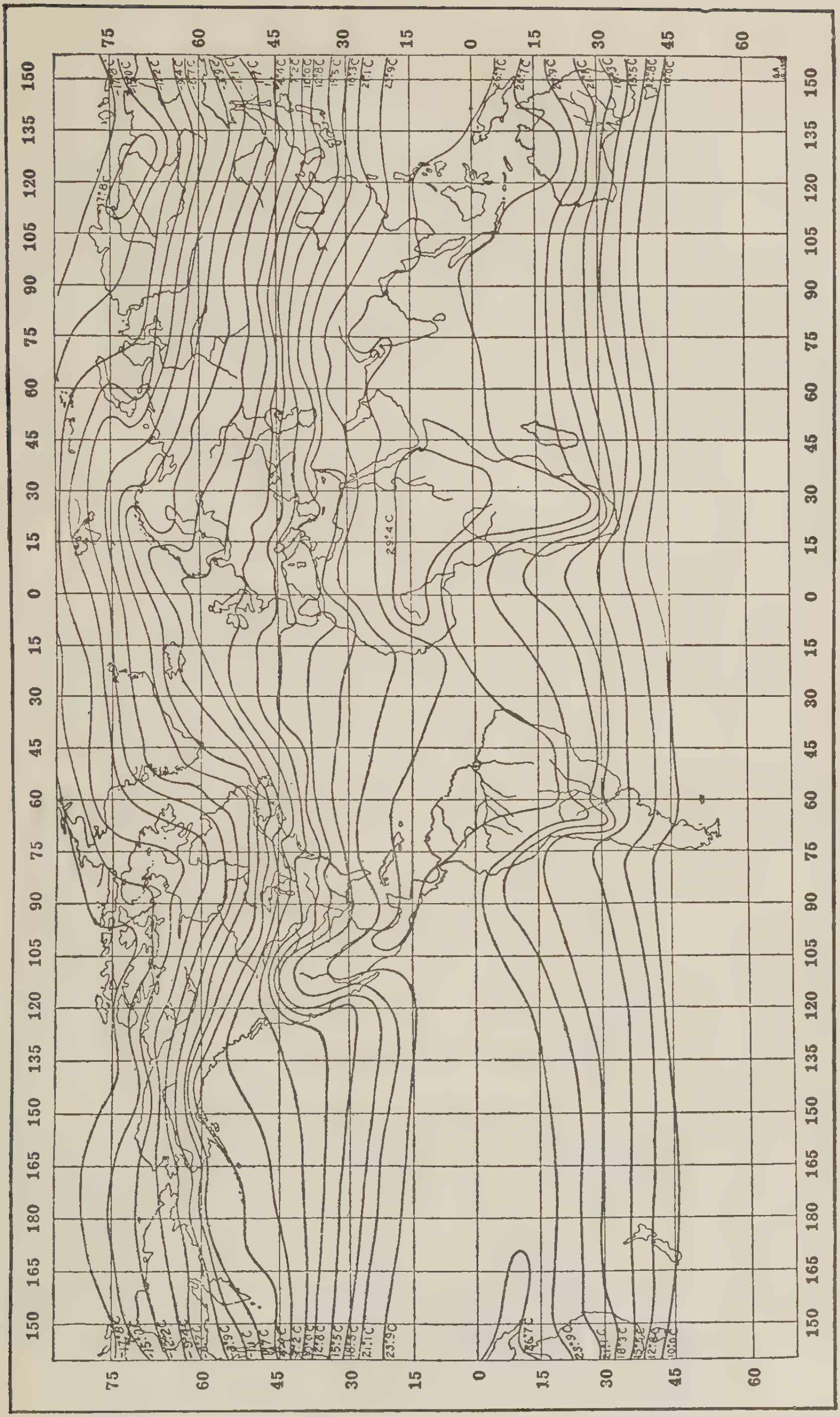

FIGURE 33. Average annual isotherms, °C (after Buchan).

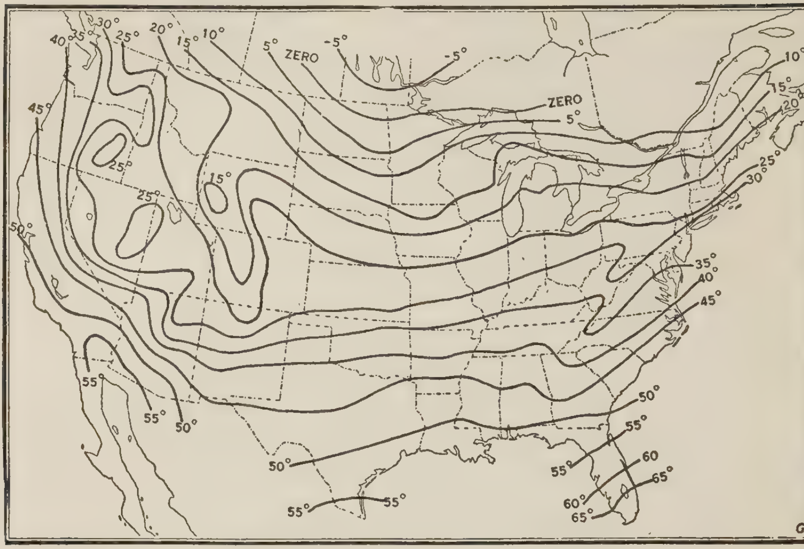

FIGURE 34. Average January temperatures, °F, in the United States.

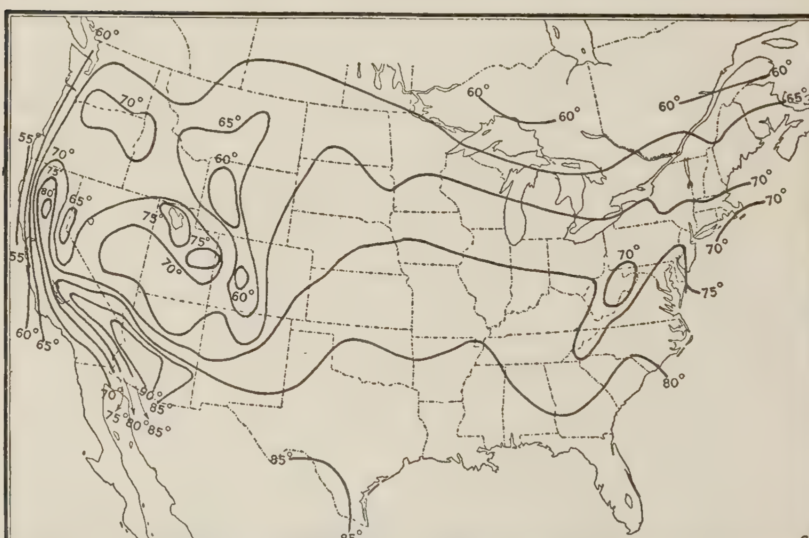

FIGURE 35. Average July temperatures, °F, in the United States.

Annual Range in the United States.—Figures 34 and 35 give mean temperatures in January and July for the United States. The isotherms bend southward in the interior during the winter and northward during the summer. Near the coasts the annual range is much less. The effects of irregular topography are well shown in the western part of the country and, to a smaller extent, in the region of the Appalachian Mountains. In most parts of the northern hemisphere the coldest month is January and the hottest month is July, or, in other words, the temperature lags behind the times of greatest and least heat received by the sun. This is due to the cumulative effect of heat or cold. In the early summer, the atmosphere is still cool from the effects of the preceding winter and in northern regions snow covers the ground until late spring. Much of the sun's heat is required to overcome these influences. When this has been accomplished, the effect of insolation increases for a time, even though the amount received is less than before. In the winter the opposite of this occurs.

Diurnal Range.—Typical illustrations of this may be seen in the lowest curves in Figures 26 and 27, pages 40 and 41. The hottest part of the day occurs between 2 and 4 P. M., when the amount of incoming radiation is just balanced by that of outgoing radiation, a case analogous to the one discussed in the preceding paragraph. The coldest part of the day occurs just before sunrise, or just before insolation becomes equal to terrestrial radiation. These curves show conditions on relatively clear days. The range is greater when absolutely clear weather prevails, especially if the relative humidity is low, and less on cloudy days. When the sky is overcast with low dense stratus clouds, particularly if rain or snow is falling, the diurnal range may be less than one degree and occasionally may be entirely masked by storm conditions, the maximum occurring at night and the minimum during the day. In the interior of continents, where humidity is low, much larger ranges occur, and near the oceans, where humidity is high and where, moreover, the tempering effect of the water itself is felt, relatively small ranges occur, especially if further modified by land and sea breezes. Other local influences such as mountain and valley breezes, chinooks, thunderstorms, etc., may at times and often do produce irregularities in the daily march of temperature, but, on the average, in spite of all these departures from normal conditions, the average range for any place follows the general trend of those in the figures.

CHAPTER IV.

ATMOSPHERIC PRESSURE.

General Statement.—Of all the meteorological elements, atmospheric pressure probably exerts the least noticeable effect upon the average person. We are immediately and often painfully aware of changes in temperature, moisture and wind, but almost entirely unconscious of the ordinary changes in pressure that occur from day to day. Only when we ascend a mountain or make a flight in an airplane or a balloon do we become conscious of effects caused by the large reduction of air pressure which then occurs. Yet in many respects this element is highly important. It forces water to rise in a pump, when the pressure of the air therein has been diminished by means of a piston; its measurement makes possible the determination of altitudes; its horizontal variations give rise to winds and the consequent changes in temperature, moisture, etc., at different times and places; and its accurate observation over wide areas enables the forecasters to predict with considerable success the coming weather changes. Moreover, it has in recent years assumed peculiar significance in its relation to aviation: the aviator is limited in the height to which he can fly by its effect upon himself and upon the performance of his engine and plane. As already stated, its measurement enables him to determine approximately the altitudes which he reaches.

Early in the 17th century Galileo first observed that there is a limit beyond which it could not be said that "nature abhors a vacuum." This conclusion was based upon his discovery that water would not rise above a certain level in a pump. Torricelli carried this investigation farther and showed that the height of a column of water that is sustained by the atmosphere diminished with altitude. In 1643 he substituted mercury for water, and the result was the mercurial barometer which has since been in general use for the measurement of pressures at the earth's surface. This type of barometer is too heavy and at the same time too delicate for use in free air observations and a specially designed instrument known as the aneroid barometer has therefore been constructed and has been generally adopted for such work. (For description of barometers see Chapter II, pp. 10, 20.)

Units of Measurement.—Atmospheric pressures have been for the

most part expressed in terms of the linear height of the mercury column, corrected for temperature, gravity, etc., that is sustained by the pressure of the air at any given time and place. The units generally used are the inch and the millimeter. Thus, a standard sea level pressure at a temperature of melting ice and at standard gravity is given as 29.92 inches, or 760 millimeters. In steam engineering it is customary to express pressure in terms of force or weight, like pounds per square inch, and in these units the average air pressure at sea level is about 14.7 pounds per square inch. Recently, meteorologists in some countries have, for the sake of ready comparison with other units of force, agreed upon the name "bar" as the proper unit for expressing pressure. A bar, as defined by the International Commission for Scientific Aeronautics, is a force of 1,000,000 dynes per square centimeter. It is equivalent to the pressure of a mercury column of practically 750.1 millimeters or 29.53 inches. Hence, standard sea level pressure, 760 mm., or 29.92 inches, equals 1.0133 bars, or 1013.3 millibars, the millibar being in general use because of greater convenience in placing the decimal point. In the present discussion millimeters and inches will be used, because of the large amount of data that have not yet been reduced to units of force.

Vertical Distribution of Pressure.—As one ascends in the atmosphere less and less air remains above his level and more and more is found below him. Since all particles of the air are drawn to the earth by the attraction of gravitation the pressure at any point is proportional to the weight of the column of air above the level in question. Since, also, air is compressible the lower layers are much compressed by the superincumbent weight of the overlying masses which in turn expand and at great elevations attain conditions of extreme rarity. From these considerations it is obvious that the pressure of the air must continuously decrease with elevation above the surface of the earth. While the rate of decrease varies with changes of temperature and atmospheric moisture, nevertheless there cannot be *inversions* of pressure (that is, increases of pressure with elevation) such as we find in the case of temperature.

Hypsometric Equation.—The orderly manner in which pressure decreases with elevation can be represented with great accuracy by means of a somewhat complex formula, known as the hypsometric equation (p. 61). This equation of relation between the pressure, temperature, moisture and some other conditions of an air column and the elevation above sea of the point at which the pressure is observed, is of the greatest practical importance. It must be used in calculating

the height of mountains, kites, balloons, airplanes and in all such cases where elevation is calculated from observations with barometers, barographs, altimeters, etc.

The rate of decrease is according to a logarithmic law which is too complex to discuss here but which has been set out in more or less elaborate tables which the observer must use when he wishes to calculate accurately altitudes from observed barometric pressures.

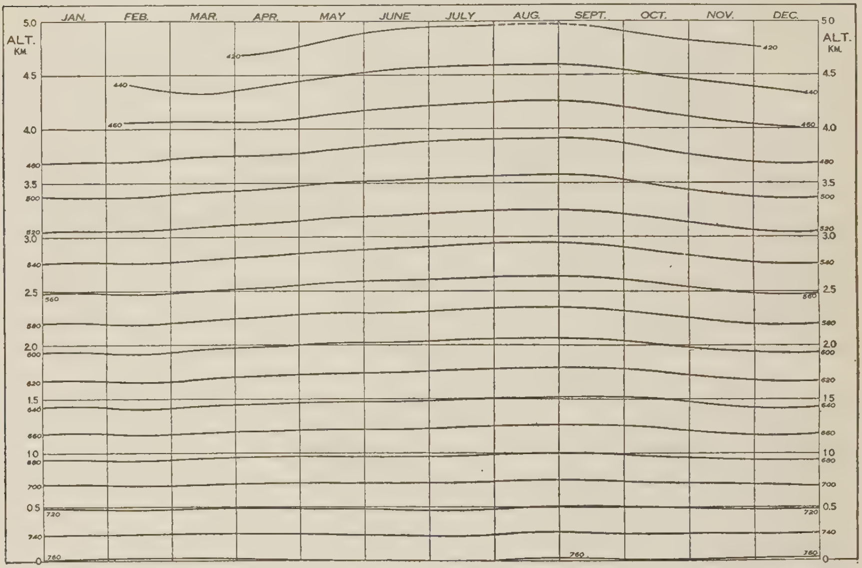

FIGURE 36. Mean annual march of free air pressures, mm., above Mount Weather, Va. (Extended to sea level.)

Altimeters, which are now regularly manufactured and form part of the equipment of airplanes, are nothing more than barometers with scales reading in feet or meters of altitude instead of pressure. The thoughtful student will notice, however, that, since the change of pressure with altitude depends very much upon the temperature, and in a lesser degree upon the moisture and other properties of a given air column, it is plain that the scale of an altimeter, however perfect the instrument may be otherwise, cannot give correct indications except for some particular mean temperature and humidity of the air column. Some important corrections are therefore required to the indications of altimeters when the temperature conditions are markedly different from those assumed in fixing the graduations of the instruments.

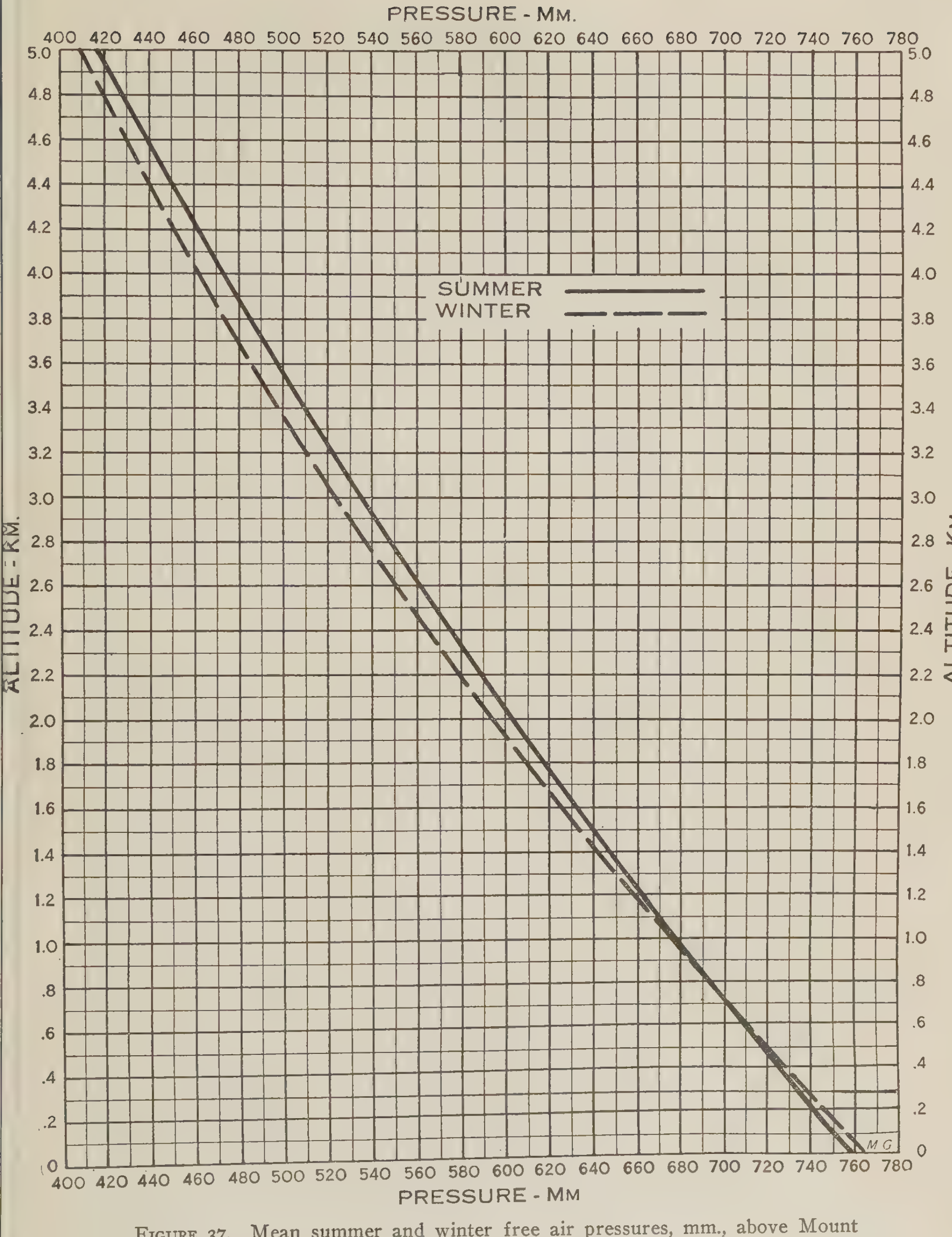

FIGURE 37. Mean summer and winter free air pressures, mm., above Mount Weather, Va. (Extended to sea level.)

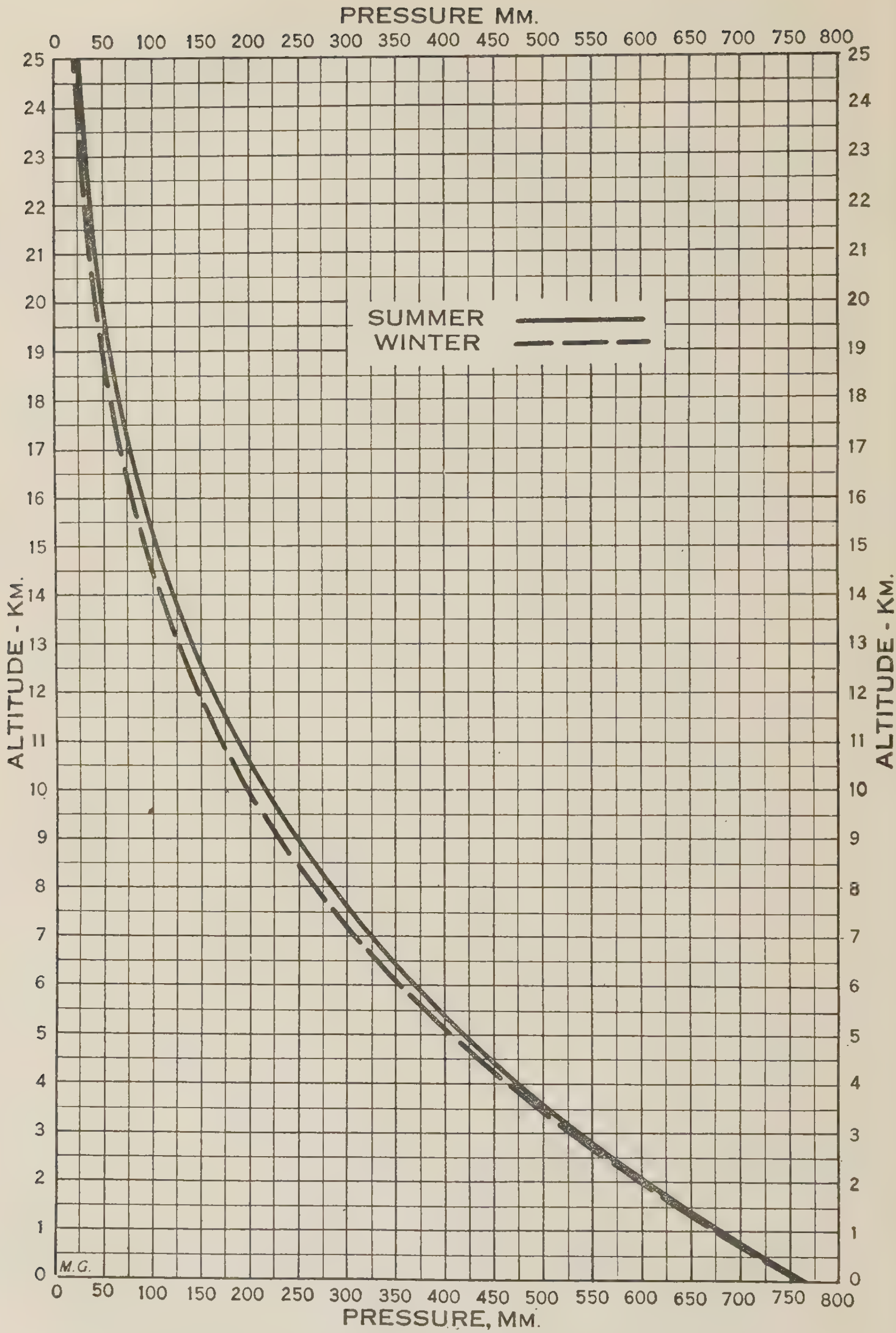

FIGURE 38. Mean summer and winter free air pressures, mm., from sounding balloon records in the Central and Western States.

With the same pressure at the earth's surface, the pressure at any height above it will vary directly with the temperature, i. e., it will be higher than normal when the mean temperature of the air column is above normal and lower when the latter is below normal. This is because cold air is contracted and a greater proportion of the air column is found in the lower levels; conversely at high temperatures the density diminishes, the air expands and a greater proportion of it is therefore found at the upper levels. These relations are well shown in Figure 36, based upon observations made at Mount Weather, Va. Mean pressures at the earth's surface do not vary greatly throughout the year, being somewhat higher in winter than in summer. At a definite elevation in the free air, however, pressures are higher in the summer than in the winter, because the expansion of the air with its higher temperature of summer lifts more of the atmosphere above the particular level in the summer time, hence increases the pressure at that level. This seasonal difference is well illustrated in Figures 37 and 38, based respectively on observations with kites at Mount Weather and by means of sounding balloons at various points in the United States. In the stratosphere the variation decreases because of the smaller differences in temperature in the two seasons. (See Chapter III.)

Vertical Distribution of Density.—A knowledge of the density of the air is of great practical importance in problems connected with aeronautics, the flight of large projectiles and the study of atmospheric phenomena. This density varies directly with the pressure of the air and inversely with the temperature and moisture content. The student should learn how to compute the density when the conditions upon which it depends are known, and as the formula is a simple one it is given as follows:

In metric units the density D in kilograms per cubic meter is

$$D = 0.46446 \frac{B - .378E}{273 + t} \qquad (1)$$

in which B = the barometric pressure in mm.
E = vapor pressure in mm.
t = temperature of the air in Centigrade degrees.

For pressures in inches and temperature in Fahrenheit degrees the same formula becomes

$$D = 1.3245 \frac{B - .378E}{459 + t} \qquad (2)$$

And D is then the density in pounds per cubic feet.

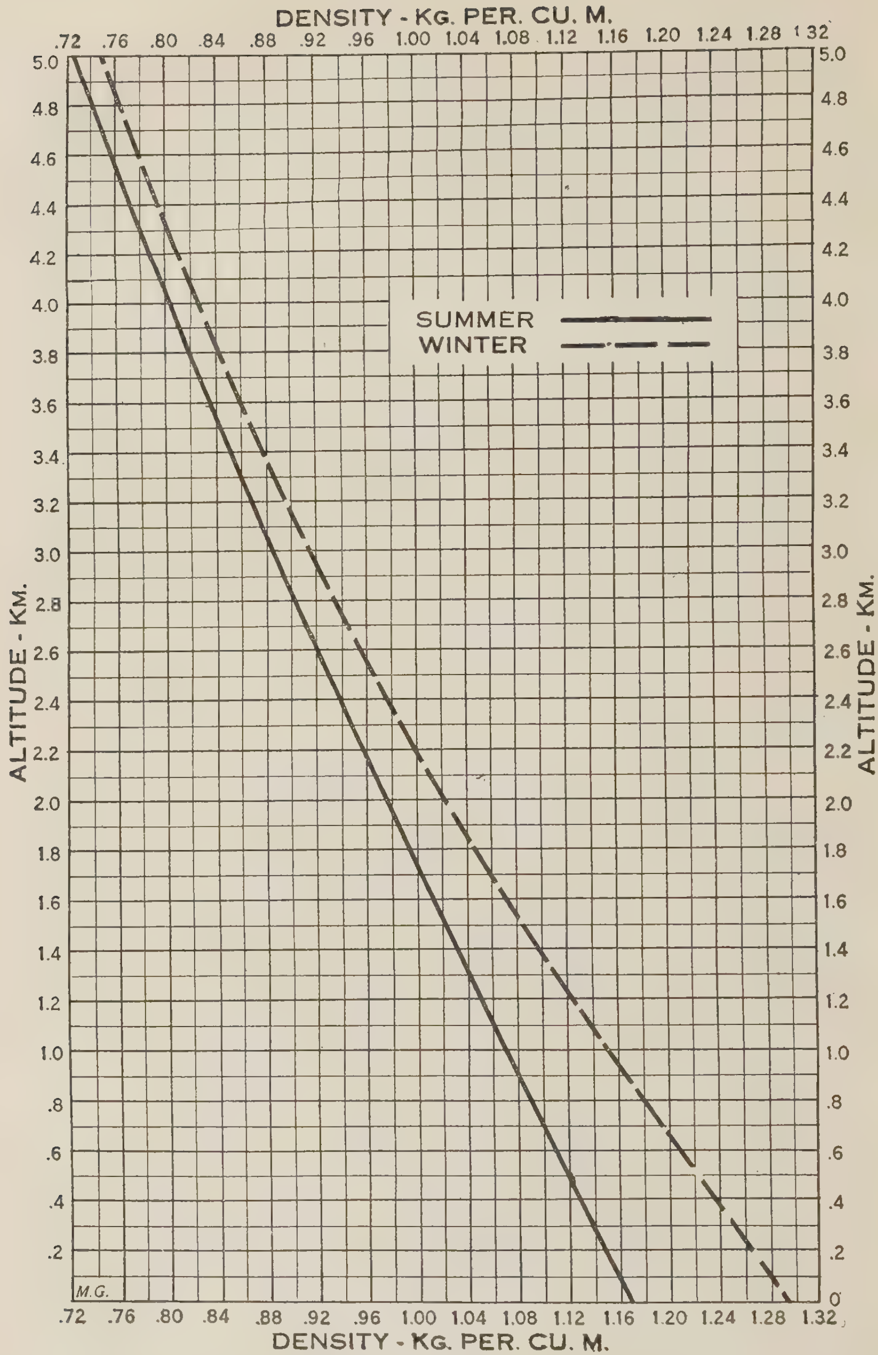

FIGURE 39. Mean summer and winter free air densities, Kg. per cu.m., above Mount Weather, Va. (Extended to sea level.)

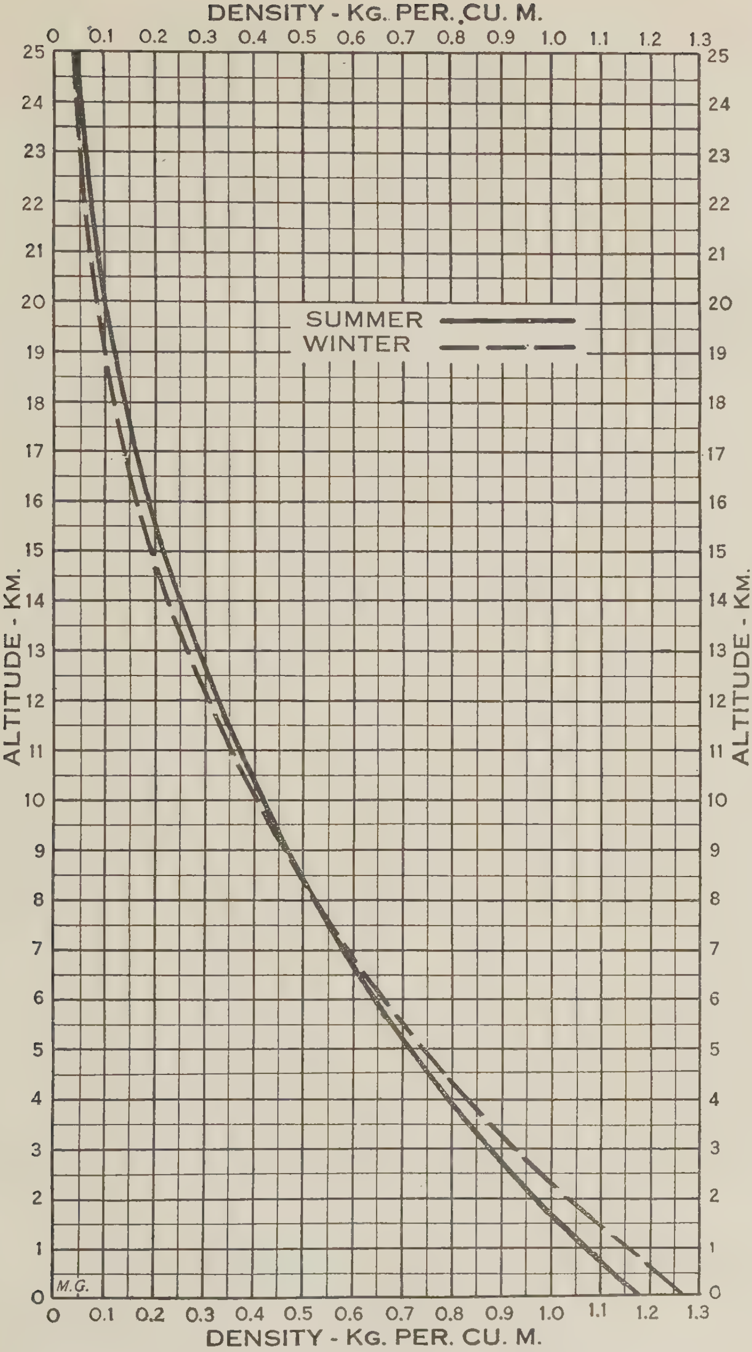

FIGURE 40. Mean summer and winter free air densities, Kg. per cu.m., from sounding balloon records in the Central and Western States.

Equations (1) and (2) enable us to compute the density for any conditions of pressure, temperature and moisture content.

Densities are subject to large changes from day to day, particularly at the surface, as is shown in Figure 39, from which it appears that the normal annual range at Mount Weather is about 10 per cent. The range from one day to the next may easily be as large as this, owing to changes in pressure and temperature, and the extreme range for the entire year may be 20 per cent. or more. It is higher in the temperate than in the torrid or arctic zones, higher at lowland than at mountain stations and much higher over the land than over the oceans.

The range is much less above the earth's surface, however, for, as previously stated, higher pressures accompany and in fact are the result of higher temperatures at those altitudes and the two, to some extent, offset each other in their effect upon the densities. Indeed, at great heights, as indicated in Figure 40, the annual range in pressure is sufficiently large to overcome the temperature effect, with the result that above 9 kilometers the mean densities are higher in summer than in winter, the exact opposite of that which occurs in the region below 9 kilometers.

Horizontal Distribution of Pressure in Relation to Temperature.—We have already seen that temperature exerts a marked influence on the vertical distribution of pressure. (See Figures 36, 37 and 38.) Because of this influence temperature is one of the most important factors also in the horizontal distribution of pressure. For, whenever and wherever the temperature is high, the entire atmosphere is elevated and the upper portions necessarily flow out over regions which are colder, thus diminishing the surface pressure over the former places and increasing them over the latter. Thus, one would at first thought suppose that pressures in the equatorial regions would be lowest and that they would increase to a maximum over the poles. That is what would actually occur on a non-rotating earth, but, as will be shown in Chapter VIII, this rotation causes the overflowing air from the tropics to turn to the right in the northern hemisphere, to the left in the southern hemisphere, thus setting up a circumpolar whirl and tending to produce very low pressures at the poles themselves. The ultimate result of these two influences is a distribution as follows: A belt of low pressure in the tropics; belts of high pressure at latitude 30° N. and S.; belts of low pressure at latitudes 60° N. and S.; and slightly increasing pressure from latitude 60° toward the poles. This planetary distribution of pressure is well shown in Figure 41.

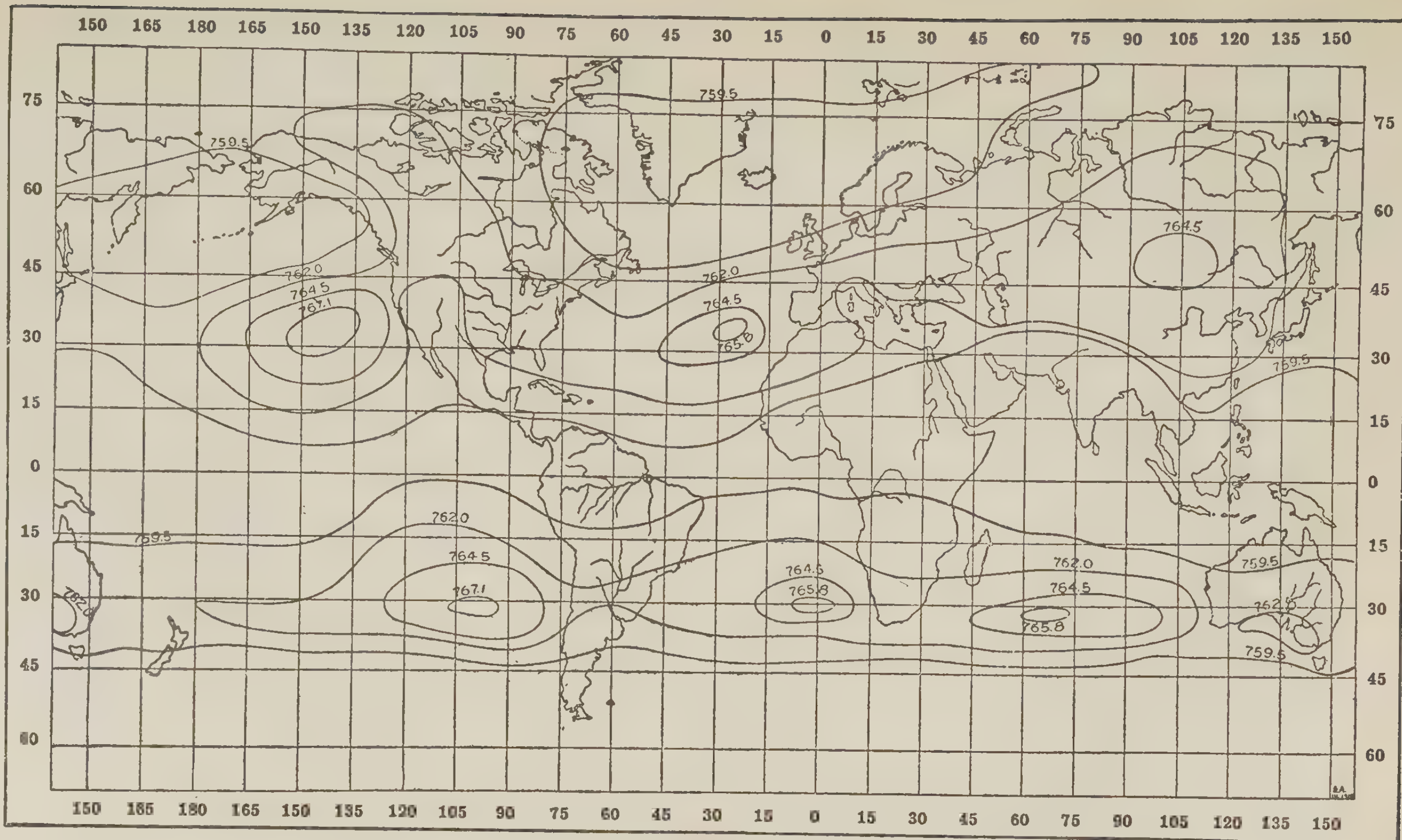

FIGURE 41. Average annual sea level isobars, mm. (After Buchan.)

Effects of Land and Water Surfaces on Annual Range.—It will be observed from this chart that the belts are not perfectly continuous, but are interrupted by the continents. This again is purely a temperature effect, the land masses having higher temperature on the average than the oceans. Their influence is better illustrated by Figures 42 and 43 which give, respectively, the average pressures over the United States for January and July. In the interior they are high during winter and low during summer, because of the seasonal change in temperature, as compared with that prevailing over the oceans. The result is that (confining our attention to the northern hemisphere) we find, in winter, high pressure over North America and Asia, with a moderate extension of the latter to include Europe, and low pressure over the northern Pacific and northern Atlantic. In the summer, pressures are high over the northern Pacific and central North Atlantic and low over North America and Asia. The annual range is much larger over the continents than over the oceans, because of the greater extremes in temperature. The difference would be much greater than it is were it not for the fact that the viscosity of the atmosphere is too small to enable it to maintain any considerable pressure gradient. Summarizing, briefly, it may be said that the annual range is greater in the temperate zones than in the tropics or the frigid zones, and greater over the continents than over the oceans, and that these results are due in large part to the unequal heating of land and water surfaces.

Diurnal and Semi-diurnal Pressure Changes.—Humphreys in Physics of the Air, Chapter XI, gives the following discussion of these daily variations: "There are two classes of well-defined 24-hour pressure changes. One obtains at places of considerable elevation and is marked by a barometric maximum during the warmest hours and minimum during the coldest. The other applies to low, especially sea level, stations, and is the reverse of the above, the maximum occurring during the coldest hours and the minimum during the warmest.

"The first class of changes just mentioned, the one that concerns elevated stations, is due essentially to volume expansion and contraction of the atmosphere caused by heating and cooling respectively. Thus the lower atmosphere over that side of the earth which is exposed to insolation becomes more or less heated, and therefore, because of the resulting expansion, its center of mass is correspondingly raised. Conversely, during the night the atmosphere cools and contracts and the center of mass is proportionately lowered. Hence, so far as this effect alone is concerned, a mountain station 1000 meters, say, above sea level, will have the greatest mass of air above it when

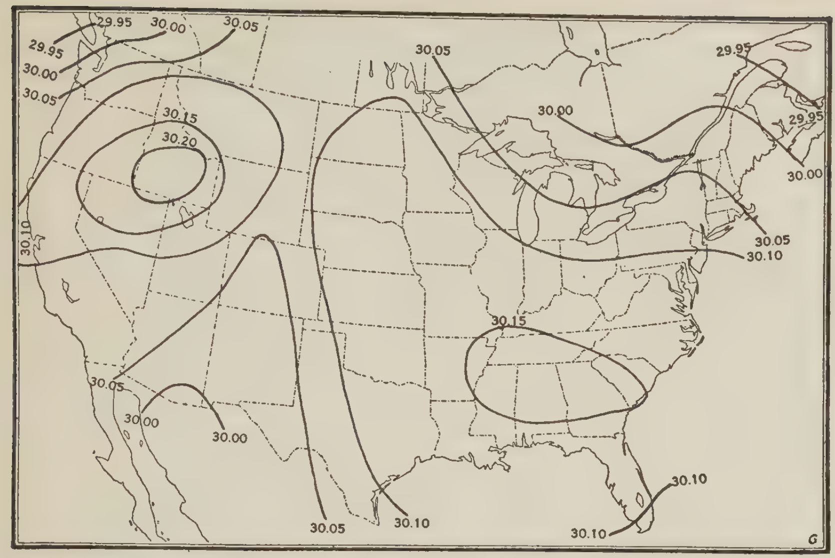

FIGURE 42. Average January sea level pressures, inches, in the United States.

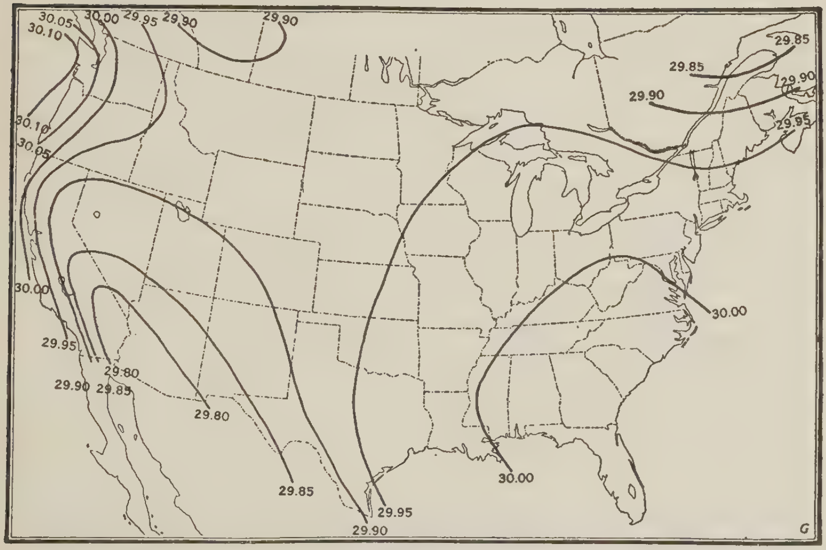

FIGURE 43. Average July sea level pressures, inches, in the United States.

the atmosphere below is warmest or most expanded, and the least when the lower atmosphere is coldest or most contracted—that is to say, this effect tends to produce, at such stations, barometric maxima during afternoons and minima about dawn.

"There is, however, another effect resulting from the volume expansion and contraction of the atmosphere to consider; namely, its lateral flow. To this, mainly, is due that daily barometric swing at sea level, the early evening minimum and the early morning maximum, that is the reverse of the high-level oscillation. The expansion and consequent vertical rise of the air on the warming side of the earth together with the simultaneous contraction and fall of the atmosphere on the cooling side, establishes a pressure gradient at *all* levels of the atmosphere directed from the warmer toward the cooler regions, a gradient that obviously causes the well-known heliotropic wind—the wind that turns with the sun—and thus leads to maximum pressures at the coldest places and minimum pressures at the warmest. But as these regions are along meridians, roughly, 10 hours, or 150 degrees apart, and perpetually move around the earth at the rate of one revolution every 24 hours, there must be a corresponding perpetual flow of air, or change of flow, as above described, in a ceaseless effort to establish an equilibrium which, since the disturbance is continuous, can never be attained.

"Barometric records, when averaged for a period of several years, show conspicuous 12-hour cyclic changes that culminate in maxima, and minima, at approximately 10 o'clock A. M. and P. M., and 4 o'clock A. M. and P. M., respectively—the exact hour depending somewhat upon season, elevation, and, presumably, weather conditions. This phenomenon is well illustrated by Figure 44, which gives, from hourly values, the actual average daily pressure curve for each month, and also for the entire year, as observed at Key West, Fla., during the years 1891-1904.

"Some of the observed facts in regard to this 12-hour cyclic change of pressure are:

"(a) The amplitude is greatest in the tropics and decreases toward the poles, approximately as the square of the cosine of the latitude.

"(b) The amplitude is everywhere greatest on equinoxes and everywhere least on solstices.

"(c) The amplitude is greater at perihelion than at aphelion.

"(d) The amplitude is greater by day than by night.

"(e) The amplitude is greater on clear days and least on cloudy.

"(f) The day amplitude is greater over land than over water.

"(g) The night amplitude is greater over oceans than over continents.

"(h) Over the tropical Pacific Ocean the forenoon barometric

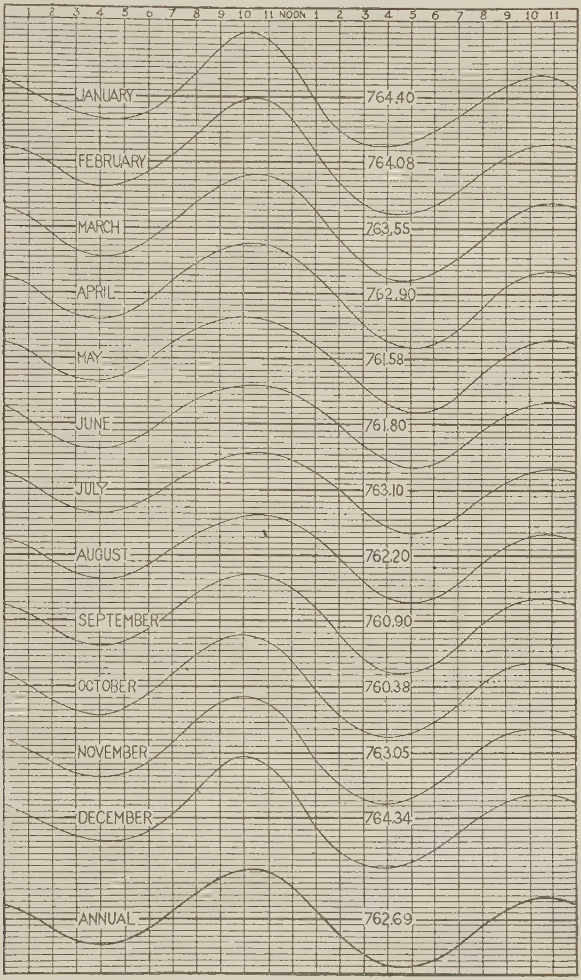

FIGURE 44. Average daily barometric curves, mm., Key West, Fla.

maximum is about 1 mm. above and the afternoon minimum 1 mm. below the general average pressure.

"Obviously, other things being equal, both the daily change in temperature and the resulting change in convection are greater in the

tropics than elsewhere; greater at perihelion than aphelion; greater during clear weather than cloudy; greater over land than over water; and greatest when the time of heating and the time of cooling (day and night) are equal, and least when these are most unequal or at the times of solstice. Hence all the above facts of observation strongly favor, if they do not compel, the conclusion that the daily cyclic pressure changes are somehow results of daily temperature changes. There are, however, a number of other causes of slight pressure changes, but apparently only the following have any appreciable value:

"1. Horizontal flow of the atmosphere from the regions where it is most expanded toward those where it is most contracted.

"2. Interference by vertical convection with free horizontal flow.

"3. Natural or free vibration of the atmosphere as a whole."

From a detailed study of these three factors, the following conclusions are reached:

"The course of events at each locality appears to be substantially as follows:

"1. A forced afternoon compression of the atmosphere, followed by its equally forced afternoon expansion, the two together forming one complete barometric wave, with a 10 o'clock maximum and a 4 o'clock minimum, in harmony with the free vibration of the entire atmospheric shell.

"2. Nondisturbance through the night or during the time of a single free vibration.

"3. Repetition the following day of the forced disturbances in synchronism with, and therefore at such time as to reënforce, the free vibrations.

"The series of disturbances is continuous, forced by day and free by night, but the resulting amplitudes of the barometric changes are limited, through friction and through the absence of perfect synchronism, to comparatively small values. Each point upon the atmospheric shell receives at every alternate swing a forced impulse in phase with the free vibration, and therefore at such time and in such manner as indefinitely to maintain the vibrations of the atmosphere as a whole.

"The forenoon maximum and the afternoon minimum are primary disturbances equally forced but in different ways by the daily increase of temperature, while the evening maximum and the morning minimum are secondary disturbances caused by the joint action of the forced

primaries through the 12-hour free vibration of the atmosphere. In short, the semi-diurnal swing of the barometer as a result of merely fortuitous circumstances—of the fact that the mass of the atmosphere happens to be such that the period of its free vibration is approximately just one-half that of the earth's rotation."

Irregular Pressure Changes.—The most important cause of these irregular variations is the alternate passing of areas of high and low pressure. In the temperate zones they occur at all times of the year, but are best developed and most frequent in the winter. (See Chapters IX and X for a full discussion of their characteristics, movements, etc.) Figure 45 illustrates the abrupt pressure changes that they may

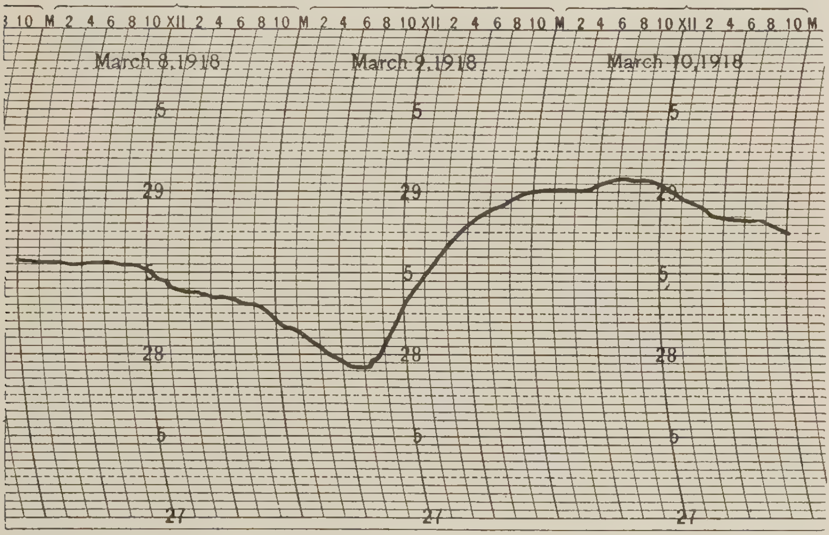

FIGURE 45. Pressure changes, inches, at Drexel, Nebr., March 8 to 10, 1918, inclusive, due to an intense low followed by a well developed high.

bring to a station over which they pass. In the tropics storms of great intensity, commonly known as hurricanes, produce very low pressures, but these are of rare occurrence, or at any rate the frequency with which they visit any particular locality is small. The relative effect of highs and lows upon the mean pressure range at different latitudes in the northern hemisphere for January and July (excluding storms of exceptional severity) is shown in the following table, copied from the Barometer Manual of the British Meteorological Office.

Latitude N.	Mean Range. January mm.	July mm.
0° to 23½°	5–10	5–7.5
23½° to 30°	10–16.5	7.5–10
30° to 40°	16.5–31.5	10–15
40° to 50°	31.5–38	15–20
50° to 60°	38–46	20–25.5
60° to 65°	46–43.5	25.5

Other causes of irregular changes are tornadoes and waterspouts, which, though infrequent and of short duration, may produce abnormally low pressures; and thunderstorms, which often occur in certain regions and places and usually cause an abrupt change, usually a rise, in pressure. The amount of this change is, however, small and does not materially affect mean values.

CHAPTER V.

EVAPORATION AND CONDENSATION.

INTRODUCTION.

The presence of water vapor in the atmosphere is of such vital importance in the economy of Nature, and the source of so many phenomena, as to demand a study of, among other things: evaporation, by which the vapor is gotten into and rendered a portion of the atmosphere, mainly from free surfaces, but also from vegetation and damp soil; and condensation, by which in various forms it is removed from the air.

EVAPORATION.

Evaporation, the process by which a liquid becomes a vapor, or gas, is a result of the kinetic energy of the individual molecules. Some of the molecules at or near the surface have such velocities and directions that they escape from the liquid and thus become an integral part of the surrounding gas or atmosphere; and as the chance of escape, other things remaining equal, increases with the velocity, it follows (*a*) that the average kinetic energy of the escaping molecules is greater than that of the remaining ones, or that evaporation decreases the temperature of a liquid, and (*b*) that the rate of evaporation increases with increase of temperature.

Just as the kinetic energy of some of the molecules of the liquid carries them into the space about, so, too, the kinetic energy of some of the molecules of the gaseous phase causes them to penetrate into and thus become a part of the liquid. In reality, therefore, evaporation from and condensation onto the surface of a liquid, though necessarily taking place by discrete molecular units, practically are continuous processes whose ratio may have any value whatever. As popularly used, however, and even as very commonly used scientifically, the term "evaporation" refers to the net loss of a liquid, and "condensation" to its net gain, so that, in this sense, both are said to be zero when, as a matter of fact, they are only equal to each other.

In the sense of net loss, which admits of accurate measurement, evaporation has been the subject of numerous investigations. Vege-

tation, soil, and the free water surface each offers its own peculiar and numerous evaporation problems. In what follows, however, only the free surface will be considered.

Evaporation Into Still Air.—The problem of evaporation into absolutely still air has been definitely solved by Stefan,[1] in special limited cases. The discussions, however, are necessarily tedious, and in the end have but little bearing on the evaporation problems that confront the meteorologist and the engineer, who are concerned with evaporation as it occurs under the variable conditions of out-doors, rather than under the fixed and radically different conditions of the laboratory.

Evaporation in the Open.—Several hundred papers,[2] many of them giving the results of elaborate investigations, have been published on the evaporation of water from free surfaces, vegetation, and soil, and, while no equation has been found that expresses in terms of easily measurable quantities the rates of evaporation in the open, nevertheless several factors that control these rates have been discovered and more or less approximately evaluated. In the case of free, clean surfaces the principal factors are:

(a) *Salinity.*—It has repeatedly been observed that the evaporation of salt solutions decreases with increase of concentration, and that sea-water evaporates approximately 5 per cent. less rapidly than fresh water under the same conditions.

(b) *Dryness of the Air.*—Many observations have shown that, to at least a first approximation, the rate of evaporation is directly proportional, other things being equal, to the difference in temperature indicated by the wet and dry bulb thermometers of a whirled psychrometer.

(c) *Velocity of the Wind.*—All observers agree that evaporation rapidly increases with wind velocity, but until recently only empirical equations have been developed that involve this important factor. Jeffreys,[3] however, has shown that in the case of strictly horizontal winds the rate of evaporation varies as the square root of the wind velocity, and as the three-half (1.5) power of the diameter of the evaporating vessel (assuming it to be circular).

But even this great theoretical advance does not fully solve the

[1] Sitzungsberichte der K. Akad. der Wis. Wien, 68 (1873), 385-423; and 73 (1881), 943-954.

[2] Livingston, An Annotated Bibliography of Evaporation, Monthly Weather Review, Wash., June, September, and November, 1908, and February, March, April, May, and June, 1909.

[3] Phil. Mag. 35, p. 270, 1918.

problem of the effect of wind velocity on evaporation, since in the open the wind has a variable but important vertical component, the effect of which on the rate of evaporation must be very different from that of a strictly horizontal component.

(d) *Barometric Pressure.*—Since the pressure of any gas retards the diffusion of other gas molecules, whether of the same or different nature, it follows that when the vapor tension is comparatively small evaporation must vary inversely, nearly, as the total barometric pressure.

(e) *Area of Surface.*—Obviously the total amount of water evaporated must increase with the area of the evaporating surface, other conditions being alike, but not at the same rate. If the evaporation is from a circular area into absolutely still air it increases as the square root of the area. If the evaporation is into a strictly horizontal wind it should, according to Jeffreys' equations, vary substantially as the three-fourth power of the area. However, under ordinary outdoor conditions the rate of total evaporation appears to be much more nearly, though by no means exactly, proportional to the first power of the surface.

(f) *Temperature of the Water.*—Evaporation increases rapidly with the temperature of the water, roughly in proportion to the saturation pressure at that temperature, provided the general humidity of the air is low. When, however, the water surface is colder than the dew-point temperature of the air the evaporation becomes negative; that is, condensation obtains. When the air is colder than the water surface, evaporation may continue into it after saturation has been reached and thereby produce fog, the process being one of distillation and condensation.

Even when the water is frozen, it still continues slowly to evaporate (sublime) whenever the air is sufficiently dry, but the laws governing this sublimation are not well known.

CONDENSATION.

Condensation, the process by which a vapor is reduced to a liquid or solid, is induced by: (*a*) reduction of temperature, volume remaining constant; (*b*) reduction of volume, temperature remaining constant; (*c*) a combination of temperature and volume changes that jointly reduce the total vapor capacity. In the open, water vapor is condensed: (1) by contact cooling; (2) by the mixture of masses of air of unequal temperatures; (3) by expansional or dynamic cooling

due to vertical convection, or, occasionally, other causes, especially rotation, as in tornado and waterspout funnels.

Condensation Due to Contact Cooling.—During clear nights the surface of the earth, including vegetation and other objects, loses much heat by radiation, and thus both it and the air in contact with it are reduced to lower temperatures, obviously more pronounced the gentler the winds. After the dew-point has been reached all further loss of heat, producing now a much smaller proportionate decrease of temperature, results in the deposition, respectively, of dew and hoar-

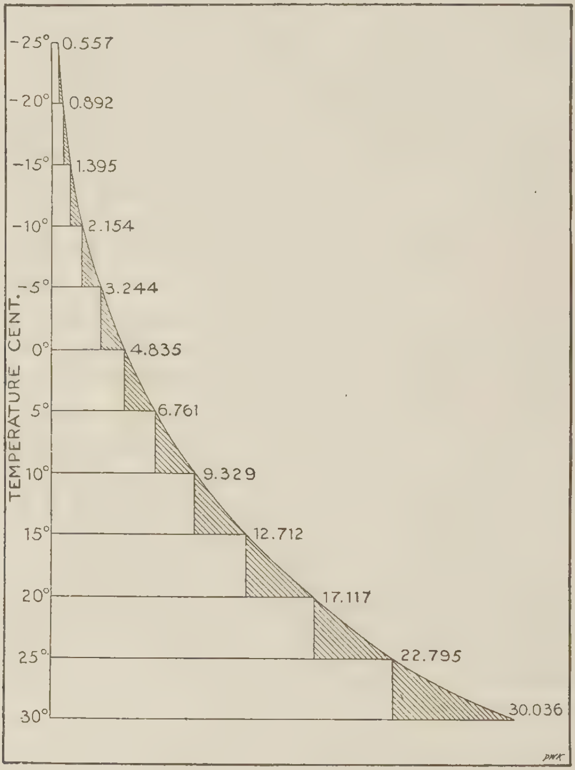

FIGURE 46. Grammes of water vapor per saturated cubic metre, at different temperatures. Bases of shaded portions proportional to precipitation per 5°C. cooling from the temperatures indicated.

frost at temperatures above and below freezing. Similarly, relatively warm, moist air moving over a snow bank, for instance, may deposit some of its moisture.

Condensation Due to Mixing.—Since the amount of water vapor per saturated unit volume decreases with temperature more rapidly than the absolute temperature itself, at least through the range of atmospheric temperatures (see Figure 46), it follows that the mixture of two saturated masses of air of unequal temperatures must produce some precipitation. The amount of precipitation induced in this manner, however, is surprisingly small; indeed, it seldom can be sufficient to produce more than a light cloud or fog. The theory involved in a full discussion of this process of producing condensation is comparatively simple, but the subject perhaps not one of great importance. Those who wish to follow it up can find it in appropriate treatises.

Condensation Due to Dynamic Cooling.—As a mass of air rises, no matter what the cause of this rise may be, it leaves more and more air below it, and thus has a less and less load of air above it to support. As the load decreases the rising mass of air expands, but obviously against whatever the load or pressure may then be. That is it does work, necessarily at the expense of its own energy—its heat. As it rises it (including, of course, such water vapor as may be present) therefore cools. But as the amount of water vapor a given space can contain under a given pressure rapidly decreases with decrease of temperature, it follows that the dynamical cooling of a rising mass of humid air is quite certain to produce condensation (cloud) at a greater or less altitude.

A complete discussion of this topic is unavoidably tedious and more or less mathematical. A condensed account with references to original papers in given in the April, 1918, number of the Journal of the Franklin Institute.

As a practical rule, however, if T_o is the centigrade temperature of the surface air and T_d the dew point, the height h at which cloud will begin to form as a result of dynamical or expansional cooling incident to convection, such as occurs in a heat thunderstorm, is given by the equation

$$h = 125\ (T_o - T_d) \text{ meters, roughly.}$$

Principal Forms of Condensation.—Condensation assumes many forms, of which the chief are: (*a*) *Free drops,* varying in size all the way from the fog or cloud particle up to the largest rain drop, or from .03 mm., roughly, to about 5 mm. in diameter. (*b*) *Dew,* water that has condensed on objects that by any process have attained a temperature below the current dew-point of the air immediately in contact with the bedewed objects. The cooling necessary to the for-

mation of dew usually is owing to loss of heat by radiation. (*c*) *Frost,* a light feathery deposit of ice caused by the same process that produces dew, but occurring when the temperatures of the objects on which it forms are below freezing. (*d*) *Rime,* a frost-like deposit of ice, often several inches deep, on the windward sides of exposed objects. It is formed from impinging undercooled fog particles, and hence grows straight into the wind. (*e*) *Glaze* (ice storm), a coating of clear smooth ice on the ground, trees, etc. It generally is caused by the falling of rain on cold (below freezing) surfaces. (*f*) *Snow,* tabular and columnar particles of ice formed in the free air at temperatures below freezing. All are hexagonal in type but of endless variety in detail—many exquisitely beautiful. (*g*) *Sleet,* ice pellets, mere frozen rain drops (or largely melted snowflakes refrozen), due to the falling of the precipitation through a cold layer of air near the surface of the earth, and that rattle when they strike a window, for instance. (*h*) Hail, lumps of ice more or less irregular in outline, and generally consisting of concentric layers of clearish ice and compact snow. As here defined, in accordance with the usage of the United States Weather Bureau, it occurs only in connection with thunderstorms and may be of any size up to that at least of a baseball, or large orange, such as fell in considerable quantities at Annapolis and other points in Maryland on June 22, 1915.[4] Indeed much larger stones have occasionally been reported, and presumably have occurred. At any rate in some instances stock in the fields have been killed by blows from hailstones of unusual size.

Other forms of precipitation that should, perhaps, be mentioned are: *graupel,* soft snow pellets; *mist,* a thin fog; and *drizzle,* a light rain of very small drops.

Why the Atmosphere Generally is Unsaturated.—It may, perhaps, seem strange that, in spite of the continuous and rapid evaporation from nearly all parts of the earth's surface, the atmosphere as a whole never becomes even approximately saturated. This condition, however, is a necessary result of vertical convection. Obviously, whatever the temperature and relative humidity of a given mass of air at any point of its convectional route, its absolute humidity is less then, in general, than when its ascent began, by the amount of rain or snow already abandoned by it. That is, on the average, air in a convection circuit descends to the earth dryer than when it previously ascended from it. In short, convection, because it induces abundant precipitation, is therefore a most efficient drying process; and because com-

[4] Fassig, Monthly Weather Review, September, 1915.

paratively little precipitation is produced in any other way, convection alone prevents the atmosphere from becoming and remaining intolerably humid.

Summer and Winter Precipitation.—Vertical convection, essential as above explained to all considerable condensation, results from three distinct causes: (*a*) superdiabatic temperature gradients, due often to surface heating; (*b*) converging winds, as in the front half of a cyclone; and (*c*) forced rise from (1) flow over land elevations and barriers of cold air, (2) underrunning of cooler winds. The first or thunderstorm type of convection causes much of the summer precipitation of temperate regions, as also nearly all the rain of the tropics, while the second or cyclonic convection produces by far the greater part of winter precipitation, except, perhaps, that which occurs along the windward sides of the most favorably situated barriers. Also, during the colder season precipitation usually occurs lower down the barrier slope and may be induced by feebler cyclones or other storms than in the warmer. This is owing in part to the fact that generally there is less difference between the actual and dew-point temperatures during winter than during summer (a condition determined by the great seasonal temperature changes of continents with reference to the ocean), and therefore a less convection required in the first case than in the second to induce condensation, and partly to the greater rate of decrease of temperature with increase of latitude while the days are short than while they are long, a condition that favors winter precipitation by causing a greater fall of temperature during the winter season than any other for a given travel of the wind on the front or rainy side of a cyclone. That is, usually a less vertical convection and a less horizontal travel of the air—a feebler storm—suffices to induce precipitation during winter than during summer.

The contrasts, then, between summer and winter precipitation are manifold. A typical case may be illustrated by the following table:

Contrast Between Summer and Winter Precipitation.

	Summer	Winter
Rain	Usually.	Often.
Snow	Never.	Frequent.
Hail (ice lumps)	Occasionally.	Never.
Sleet (frozen rain)	Never.	Occasionally.
On barrier	High.	Low, and up.
Type of storm	Thunderstorm frequently.	Cyclone.
Strength of convection	Strong generally essential.	Feebler often sufficient.
Intensity of cyclone	Decided usually essential.	Slight often sufficient.

CHAPTER VI.

FOGS AND CLOUDS.

The deposition of dew, the forming of hoar-frost, and the sweating of ice pitchers, all examples of surface condensation, show that atmospheric moisture promptly condenses upon any object whose temperature is below the dew-point. Similarly, volume condensation takes place in the form of a fog or cloud of innumerable droplets, or ice spicules, throughout the body of ordinary air whenever by expansion or otherwise it is sufficiently cooled. But this is not equally true of all air. Thus, while the first considerable rapid expansion, and therefore decided volume cooling, of humid air in a receiver, if recently admitted unfiltered, is quite certain to produce a miniature cloud, subsequent expansions of the same air produce fewer and fewer such particles. If the old air is removed and unfiltered fresh air admitted, the condensations again occur as before; but if the fresh air enters through an efficient filter, such as a plug of cotton wool a few centimeters long, condensation remains as difficult as in the exhausted air. The admission, however, of a little smoke restores to the exhausted and confers upon the filtered air full powers of condensation.

Obviously, then, cloud droplets form about nuclei that cannot easily pass through mechanical filters of fine texture, and microscopic examinations of the residue left on the evaporation of these droplets have shown the nuclei to consist in large measure of dust particles, both mineral and organic. Hygroscopic gases, such as the oxides of sulphur and of nitrogen, may also act as condensation nuclei, but ordinarily there is abundant dust in the atmosphere (thousands of particles per cubic centimeter) to provide for all precipitation. It is often urged that free electrons in the air also act as nuclei about which water vapor condenses, but, as this type of condensation requires about a fourfold supersaturation, its occurrence in the open seems extremely improbable.

As stated, volume condensation may be induced in the atmosphere by any cooling process: whether by radiation, as on clear nights; mixing warmer with colder masses of air; movement of relatively warm air over cold surfaces, as in the case of winter south winds (northern hemisphere); or expansion, owing either to convection or barometric depression. But the cooling process has much to do with

FIGURE 47. Radiation fog, Loudoun Valley, Va. (A. J. Weed, photo.)

FIGURE 48. Advection fog, seen from Mount Wilson, Cal. (F. Ellerman, photo.)

FIGURE 49. Cirrus. (F. Ellerman, photo.)

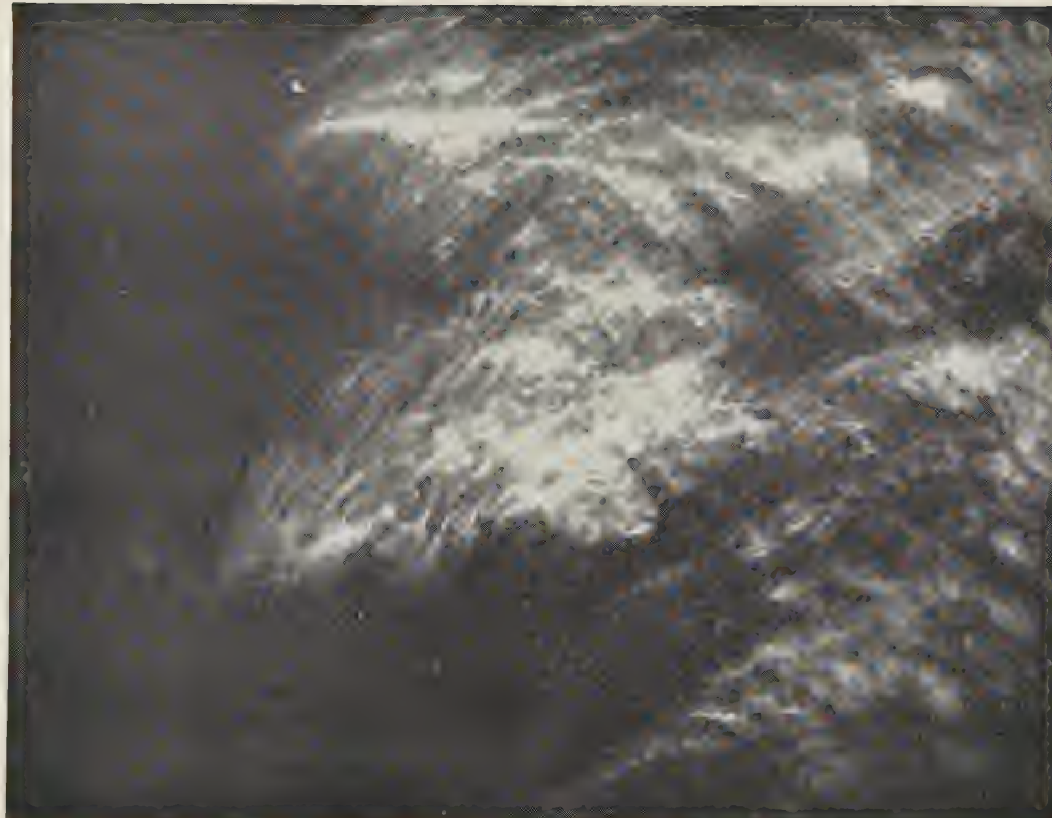

FIGURE 50. Cirrus. (F. Ellerman, photo.)

FIGURE 51. Cirro-stratus, and advection fog, seen from Mount Wilson, Cal. (F. Ellerman, photo.)

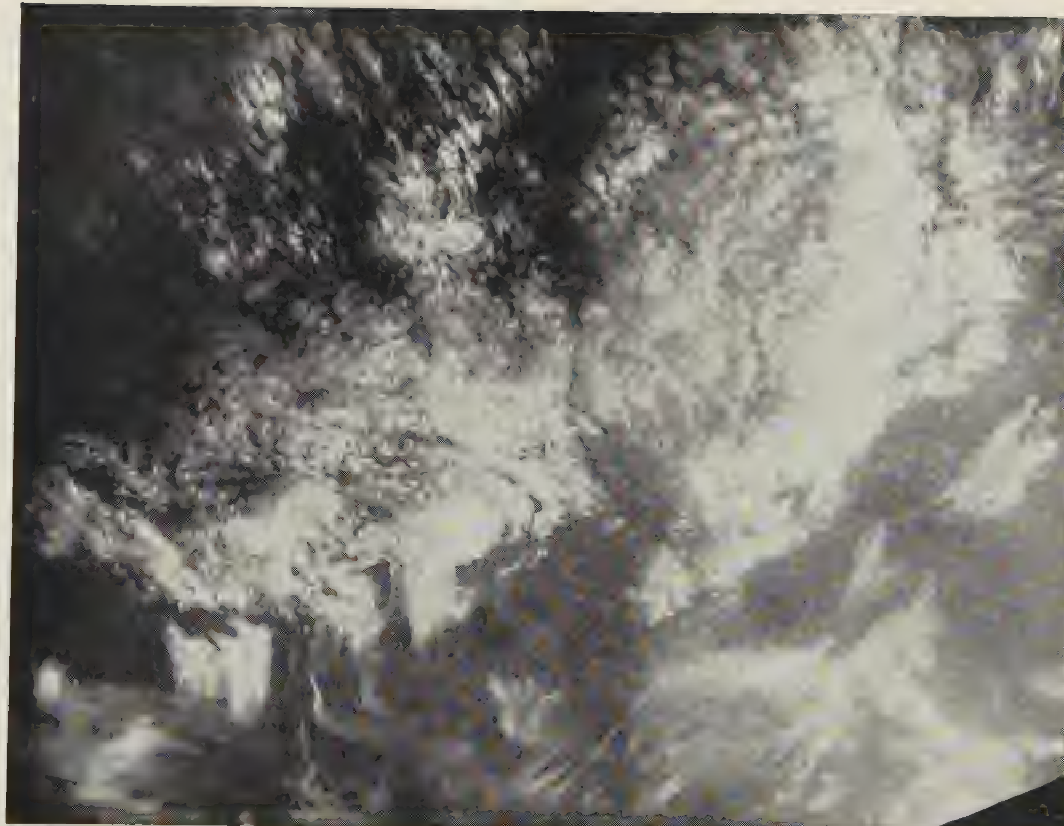

FIGURE 52. Cirro-cumuli. (F. Ellerman, photo.)

FIGURE 53. Alto-stratus and advection fog seen from Mount Wilson, Cal. (F. Ellerman, photo.)

Figure 54. Alto-cumulus. (A. J. Weed, photo.)

FIGURE 55. Strato-cumulus, or roll cumulus, near Gap Mills, W. Va. (L. W. Humphreys, photo.)

Figure 56. Cumulus, seen near Mount Wilson, Cal. (F. Ellerman, photo.)

FIGURE 57. Cumulus over island—Krakatoa. (E. E. Barnard, photo.)

FIGURE 58. Fracto-cumulus, in Monroe Co., W. Va., Peters Mountain to left. (L. W. Humphreys, photo.)

FIGURE 59. Cumulo-nimbus, over Loudoun Valley, Va. (A. J. Weed, photo.)

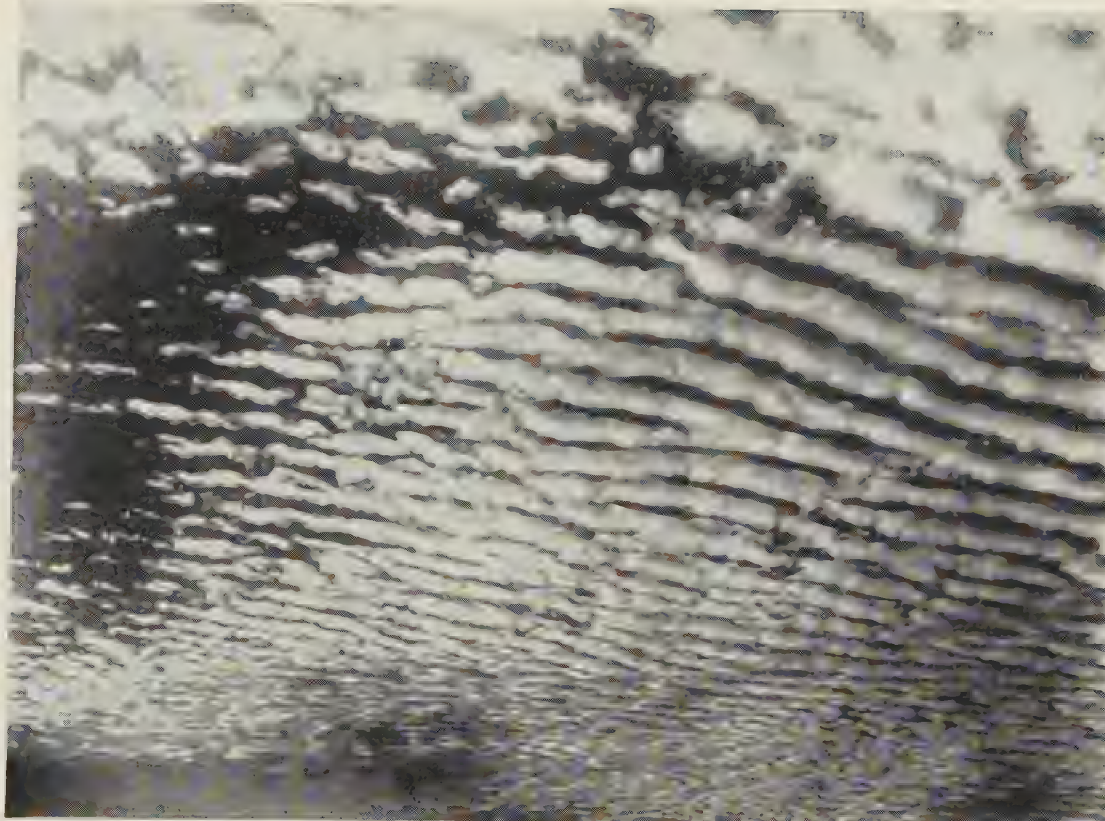

Figure 60. Billow cloud. (A. J. Henry, photo.)

FIGURE 69. Billow clouds, regular and irregular, over Washington, D. C. (A. J. Henry, photo.)

determining the extent of the condensation, the kind and amount of precipitation from it, and its general appearance, according to which, chiefly, it is classified.

Distinction Between Fog and Cloud.—Volume condensation is divided primarily into fog and cloud, but a sharp distinction between them that would enable one always to say which is which is not possible. In general, however, a fog differs from a cloud only in its location. Both are owing, as explained, to the cooling of the atmosphere to a temperature below its dew-point, but in the case of the cloud this cooling usually results from vertical convection, and hence the cloud is nearly always separated from the earth, except on mountain tops. Fog, on the other hand, is induced by relatively low temperatures at and near the surface, and commonly itself extends quite to the surface, at least during the stage of its development. In short, fog consists of water droplets or ice spicules condensed from and floating in the air near the surface; cloud, of water droplets or ice spicules condensed from and floating in the air well above the surface. Fog is a cloud on the earth; cloud a fog in the sky.

FOG.

According to the conditions under which they are formed, fogs may be divided into two general classes—radiation fogs and advection fogs.

Radiation Fog.—Fog is likely to form along rivers and creeks and even in cleared mountain valleys during any still, cloudless night of summer and, especially, autumn. In the course of a calm warm day much water is evaporated into the lower atmosphere of such regions, where in large part it remains as long as there are no winds. Hence this air, because it is humid, and the adjacent surface of the earth lose much heat during the night by radiation to the clear sky. In many cases they cool in the end to a temperature below the dew-point, and thus induce a greater or less volume condensation on the always-present dust motes that results in a correspondingly dense fog (Figure 47). Such fog, however, is not likely to occur during cloudy nights, because the air seldom then cools sufficiently, nor during high winds, since they dissipate the humidity and also through turbulence prevent the formation of excessively cold aerial lakes.

The distinctive factor in the formation of this type of fog is the free radiation of the ground and the lower air by which the latter is sufficiently cooled to induce condensation. Hence fogs formed in this

manner are properly termed "radiation fogs," sometimes also called "land fogs" and "summer fogs."

Advection Fog.—Whenever warm, humid air drifts over a cold surface its temperature is reduced throughout the lower turbulent layers by conduction to that surface and by mixture with remaining portions of the previous cold air and a correspondingly dense fog produced. Hence fog often occurs, during winter, in the front portion of a weak cyclone; also whenever air drifts from warm water to cold—from the Gulf Stream, for instance, to the Labrador Current; and wherever gentle ocean winds blow over snow-covered land—circumstances that justify the terms "winter fog" and "sea fog" (drifting on shore in places, and even some distance inland. Figure 48). Similarly, a cold wind drifting or spreading under and through a body of warm, humid air also produces a fog, though usually a comparatively light one. This explains the fog that frequently forms, during winter, along the front of a "high," and the thin fog that occasionally is seen over lakes on frosty autumn mornings, when the water appears to be steaming—actually evaporating into air already saturated and thus inducing condensation. It also explains the frequent occurrence of "frost smoke" on polar seas.

If the wind is strong the turbulence extends through a comparatively deep layer. Hence in the case of warm air drifting over a cold surface if the movement is rapid the total duration of contact between any portion of the air and that surface is likely to be so brief that but little cooling can take place and no fog be formed. Similarly, it usually also happens that fog does not form when the cold wind blowing over a warm, humid region is even moderately strong. Here the turbulence mixes the excessive humidity near the surface through so large a volume that saturation commonly is not produced, nor, therefore, any trace of fog.

From the above it appears that all fogs that result from the drifting of warm, humid air over cold surfaces, as also those that are produced by the flow of cold air over warm, humid regions, are but effects of temperature changes induced by the horizontal transportation of air; hence the proposed general name, "advection fog." The term advection is preferred to convection because the latter is practically restricted, in meteorological usage, to a change of level, whereas in the case under consideration only horizontal movements are concerned. The contradistinction, therefore, between "advection fog" and "convection cloud" is obvious, and, presumably, worth while.

CLOUDS.

The cooling of the atmosphere by which cloud condensation is induced is most frequently, perhaps, produced by vertical convection, either thermal or forced; often, presumably, by the mixing of winds of different temperatures; occasionally by pressure changes, elevation remaining the same; occasionally, also, by radiation; and rarely, in the case of very thin clouds, by diffusion and conduction.

Radiation, though productive of many fogs, is excluded from the list of principal cloud-forming processes for the reason that any mass of free air that cools in position as it must whenever its radiation exceeds its absorption, immediately gains in density and falls to a lower level where, when equilibrium is reached, it actually is *warmer* than it was before the cooling began, and its relative humidity therefore lower. Hence it seems that radiation could produce clouds only when equally active, or nearly so, over an extensive layer of practically saturated air. If radiation is unequally distributed it tends to evaporate clouds rather than produce them.

Classification.—It is not practical, however desirable, to classify clouds according to their causes, as in the case of fogs, for it often happens that the exact cause is not obvious. Hence other bases of classification have been adopted, especially form or appearance, activity, and position. Most, but not all, clouds belong to one or other of the four distinct types, *cirrus, stratus, cumulus, nimbus,* including their alto, fracto, and combination forms; alto-stratus, alto-cumulus; fracto-stratus, fracto-cumulus, fracto-nimbus; cirro-stratus, cirro-cumulus, strato-cumulus, cumulo-nimbus.

Cirrus (Ci.).—The name cirrus, literally a curl or ringlet, has been given to those fibrous white clouds that resemble great wisps of hair (mares' tails), giant curling plumes (feather clouds), tangled skeins, and various other things (Figures 49 and 50). These are the highest, often 10 to 12 kilometers above the earth in middle latitudes and still higher in tropical regions, the most tenuous, and among the most familiar of all clouds.

Since cirri usually run far ahead of the rainy portions of cyclonic areas, and grow denser as the storm approaches, it is obvious that they frequently result from cyclonic convections that extend nearly or quite to the stratosphere, where, and for some distance below which, the rising air is carried forward much faster than the storm center. But they also are fairly common in the midst of "highs" due, presumably, to a mechanical or bodily lifting of the upper air of these regions, or

overrunning of air in the general circulation, and, consequently, dynamical cooling not only of the stratosphere, as abundantly shown by the records of sounding balloons, but also of the topmost portion of the troposphere where cirri usually form.

It has been suggested that cirri often are caused by cooling in place by radiation, but, as already explained, this appears to be improbable for clouds so broken and discontinuous. On the contrary, however, it seems likely that through free radiation and cooling at night they often sink to lower levels, get *warmer* and evaporate. Thermal and mechanical convection, therefore, the first prevailing in tropical regions, the second, presumably, in extratropical, appear to be the only abundant causes of cirri.

The excessively low temperatures at which cirri are formed, generally —30° C. to —50° C., necessitate their being tenuous (at such temperatures there is but little water vapor to condense) and practically insure (exceptions have been reported[1]) that they shall consist of ice needles.

Cirro-stratus (Ci.-St.).—When cirrus clouds thicken, as they usually do on the approach of a cyclonic storm, they gradually merge into a broad cloud layer, having the appearance of a more or less continuous white veil of uneven and often fibrous texture (Figure 51), to which the name cirro-stratus has been given. Its altitude is nearly that of the cirrus, of which indeed it is only a dense and extensive form, though its under surface is not so high. Like its forerunner, the thinner cirrus, it also consists of ice crystals, as is evident from the various types of halos it forms about the sun and moon.

Cirro-cumulus (Ci.-Cu.).—Cirro-cumuli are small, fleecy cumulus clouds, generally 6 to 7 kilometers above the surface; that is, in the lower cirrus region. They usually occur in large numbers, producing an effect sometimes described as "curdled sky"; frequently, also, in groups and rows that remind one of the patterns (not the scales) on the backs of mackerel. Hence the expression "mackerel-back sky," commonly abbreviated to "mackerel sky" (Figure 52).

Their origin obviously is due chiefly to a single cause—local vertical convection, induced by unequal local heating. To each convective rise of the air there evidently must be an equivalent descent, and if the heating maxima are numerous the minima between must also be numerous, thus producing many rising currents, each with its small cumulus, surrounded by descending air and relatively clear sky.

[1] Simpson, Qr. Jr. Roy. Meteorol. Soc., 38 (1912), p. 291.

Alto-stratus (A.-St.).—The alto-stratus is a thick, grayish cloud veil (Figure 53), at times compact and fibrous in structure, and again thinner, like a heavy cirro-stratus, through which the sun or moon may dimly be seen. Its average elevation is about 4 kilometers. It may result from the forward running of air forced up by the convergence of winds in the storm area of a cyclone, from the flow of warmer over colder air, or by mere radiational cooling in place, of a layer of relatively humid air—humid from the evaporation of alto-cumuli, perhaps.

Alto-cumulus (A.-Cu.).—The name alto-cumulus has been given to those detached, fleecy clouds, with shaded portions (Figure 54), often occurring in closely packed groups and rows, that resemble enlarged cirro-cumuli. Their average altitude is approximately that of the alto-stratus—that is, 4 kilometers—and they presumably are formed by local convection, especially during fair, calm summer weather, when the relative humidity is low.

Strato-cumulus (St.-Cu.).—Strato-cumuli are large rolls of dark cloud more or less connected with thinner clouds which together cover nearly or quite the entire sky (Figure 55). Their bases are flat and at about the same height, generally 1.5 to 2 kilometers. They are formed by vertical convection, as is obvious from their rounded tops and flat bases at approximately the same level—the common saturation level.

Nimbus (Nb.).—The nimbus is any thick, extensive layer of formless cloud from which rain or snow is falling. The average altitude of its under-surface is of the order of 1 kilometer. It is produced chiefly by some type of forced convection: the converging of wind currents as occurs especially in front of cyclonic centers, the upward deflection of winds by either land or cold atmospheric barriers, and the under-running of warmer by colder air. In part, however, the cooling and consequent condensation often is owing to the mixing of cold air with warm, and to the transfer of warm air to a colder region, where it is cooled by contact, by mixing with cooler air, and by excess of radiation loss over radiation gain.

Fracto-nimbus (Fr.-Nb.).—The fracto-nimbus, popularly known as scud, is that low, detached cloud fragment, too thin and fog-like to produce rain, that occasionally is seen drifting rapidly beneath a heavy nimbus at an average elevation of probably not more than 100 to 300 meters. It seems to form only when there is considerable wind, and appears often to be caused by forced convection over cliffs or other obstacles.

Cumulus (Cu.).—The cumulus (Figure 56), often called "wool-pack," is a dense, detached cloud with a rapidly changing cauliflower head and flat base at the saturation level of rising air. Its illuminated portions are snow white, while the shaded parts are unusually dark. Its border is sharply defined and, when near the sun, very bright. The average altitude of the base is about 1.5 kilometers, and of the top rather more than 2 kilometers.

Cumuli are produced entirely by vertical convection induced by temperature differences. Hence they are always frequent in tropical regions, and also over continents at higher altitudes during summer. For the same reason, they occur over land most numerously of afternoons, and at sea late at night. At times rather low cumuli form a sort of coastal fringe along the locus of upward convection—that is, a short way out over the sea at night, and a few miles inland during the day—that might, perhaps, be called coast cumuli—attendants of the land breeze and the sea breeze, respectively. They often occur over reefs and islands (Figure 57), whose presence frequently is thus revealed while they themselves are still below the horizon. Occasionally they even parallel a large river on either side where there is rising air over the hills and bottoms and sinking over the cooler water. Further, since vertical convection depends only on the establishment of a proper vertical temperature gradient, it follows that cumuli may also form at high altitudes over the warmer portions of the ocean, or, indeed, wherever there is a sufficient temperature contrast between the surface and overlying air to induce strong upward currents.

Fracto-cumulus (Fr.-Cu.).—During the initial stages, especially, of their development cumuli often are small, and appear tattered and torn like detached and dissolving masses of fog (Figure 58). While in this condition such clouds are often called fracto-cumuli.

Cumulo-nimbus (Cu.-Nb.).—The cumulo-nimbus (Figure 59), a necessary accompaniment of every thunderstorm, is, as its name implies, a cumulus cloud from which rain is falling. It is very turbulent and much the deepest of all clouds, being anywhere from 1 to 4 or even 5 kilometers thick. Its times and places of occurrence and mode of formation are all the same as those of the cumulus.

Stratus (St.).—The stratus is a low, fog-like cloud of wide extent, often merging into a nimbus and again clearing away like lifted fog. Its average altitude is between 0.5 and 1 kilometer. It seems often to result from forced convection due to the underrunning of cold air, and also, perhaps, to the mixing of humid layers of different tempera-

tures. In some cases, that of the "velo" cloud, for instance, in southern California, it is only sea fog drifting over relatively warm land.

SPECIAL CLOUD FORMS.

Although it might seem that the above cloud types, including their numerous gradations and transitions, are exhaustive, there nevertheless are several occasional forms sufficiently distinct to justify individual names and special descriptions.

Billow Cloud.—Billow clouds (Figures 60 and 61), also called windrow clouds and wave clouds, occur in series of approximately regularly spaced bands, generally with intervening strips of clear sky. They usually form in the lower cirrus region—that is, at elevations of 6 to 8 kilometers—but may occur at any level from the surface—fogs are occasionally billowed—up to that of the higher cirrus. They are caused by the flow of one air stratum over another of different temperature and density and usually of different humidity.

It has been shown[2] that when two strata of air of different densities or vapor content flow over each other billows of great wave-length and often of large amplitude are generated in the same manner that winds produce ocean billows. As the series of waves progress the atmosphere involved obviously rises and falls, and therefore is subjected to alternate dynamical heating and cooling, with the maxima and minima temperatures corresponding to the troughs and crests respectively. Hence when the under layer is wholly or nearly saturated the wave crests are cloudy and the troughs clear. If, however, the humidity is not high, it is obvious that wind billows may exist without the incidental clouds.

It is interesting to note that, although the billow cloud appears to consist continuously of the same mass, it nevertheless is rapidly evaporating on the rear or descending portion of the wave and as speedily forming on the front or ascending portion.

Other special cloud forms are:

Lenticular Cloud.—A convex-lens shaped cloud, presumably of snow particles, often seen in high mountainous regions—over the Rocky Mountains, for instance.

Crest Cloud.—A cloud formed along a mountain (and usually rest-

[2] Helmholtz, Sitz. d. Akad. d. Wiss., Berlin, 1888, i, p. 646; 1889, ii, p. 761.
W. Wien, Sitz. d. Akad. d. Wiss., Berlin, 1894, ii, p. 509; 1895, i, p. 361.
A. Wegener, Beiträge Phys. d. fr. Atmos., 2 (1906), p. 55; 4 (1911), p. 23.

ing on it) by the upward deflection of moist air. Very common in trade wind regions, as on the island of Oahu.

Banner Cloud.—A small cloud attached to, and extending out with the wind from, the leeward side of a mountain peak.

Scarf Cloud.—A wispy fibrous cloud that first forms slightly above a rising thunder-head, and later mantles the shoulders or drapes the sides of the growing cumulus.

Mammato-cumulus.—An occasional accompaniment of thunderstorms, and consisting of numerous downward bulges in the underside of the spreading cumulus.

CLOUD HEIGHTS.

Relation to Humidity.—The heights of clouds have been measured by several obvious methods of triangulation, and from the data thus obtained it appears, as one might infer *a priori,* that whatever condition tends to increase the relative humidity tends also to lower the cloud levels, since the greater this humidity the less the amount of convectional cooling essential to condensation. Hence, in general, each type of cloud is lower in winter than summer, lower over humid than over desert regions, lower over oceans than continents, and lower with increase of latitude.

The table on page 89, copied from Hann's Lehrbuch der Meteorologie, gives the average summer and winter heights of clouds at places of widely different latitudes.

Levels of Maximum Cloudiness.—When the frequency of clouds is tabulated with reference to elevation, maxima and minima are found with the layers to which they obtain growing thicker with decrease of latitude. This phenomenon, as a whole, is interesting, but it will be necessary, in discussing it, to consider the different levels separately, since each has its own explanation.

Fog Level.—As already explained, fogs, whether caused by radiation or advection, are surface phenomena, seldom more than 100 to 200 meters thick. Hence the surface of the earth, because of the bogs that form upon it, is itself a level of maximum condensation or maximum "cloudiness."

Cumulus Level (a), *Foul Weather Type.*—Since the cumulus and the cyclone nimbus both are due to vertical convection—the first thermal, the second forced—it is obvious that the base of each occurs approximately at the saturation level; that is, the level at which a mass of air rising from the surface will have cooled to its dew-point.

Average Cloud Heights in Kilometers. (The abbreviations apply to cloud names on pages 83 to 88.)

Station	Ci.	Ci.-St.	Ci.-Cu.	A.-St.	A.-Cu.	St.-Cu.	Nb.	Cu.-Nb. top	Cu. Top	Cu. Base	Fr.-Cu.	St.
1. SUMMER, CHIEFLY APRIL TO SEPTEMBER.												
Bossekop, 70° N	8.32	6.61	5.35	4.65	3.42	1.34	0.98	3.96	2.16	1.32	....	0.66
Pavlovsk, 60° N	8.81	8.09	4.60	....	3.05	1.85	....	4.68	2.41	1.64	2.15	0.84
Upsala, 60° N	8.18	6.36	6.45	2.77	3.95	1.77	1.20	3.97	2.00	1.45	1.83	....
Potsdam, 52½° N	9.05	8.08	5.89	3.29	3.63	2.16	1.79	3.99	2.10	1.44	1.71	0.68
Trappes, 49° N	8.94	7.85	5.83	3.79	3.68	1.82	1.08	5.48	2.16	....	1.40	0.94
Toronto, 43½° N	10.90	8.94	8.88	4.24	3.52	2.06	....	....	1.70		....	....
Blue Hill, 42° N	9.52	10.10	6.67	6.25	3.76	1.16	1.19	9.03	2.90	1.78	....	0.51
Washington, 39° N	10.36	10.62	8.83	5.77	5.03	2.87	1.93	4.96	(2.45)	1.18	....	0.84
Allahabad, 25½° N	10.76	....	11.28	....	4.50	....	0.84	....	1.76		....	....
Manila, 14½° N	11.13	12.97	6.82	4.30	5.71	1.90	1.38	6.45	1.84		....	1.06
Batavia (year), 6° S	11.49	10.59	6.30	....	5.40	....	....	....	1.74		....	0.70
2. WINTER, CHIEFLY OCTOBER TO MARCH.												
Pavlovsk	8.74	7.09	5.98	....	3.17	1.50	....	....	1.60	1.12	....	1.00
Upsala	6.98	5.46	6.13	4.09	4.15	1.96	0.99	5.18	1.52	0.71	1.22	0.51
Potsdam	8.07	7.65	5.41	2.99	3.35	1.42	1.28	4.74	1.74	0.99	1.02	0.61
Trappes	8.51	5.85	5.63	3.82	4.27	1.61	1.05	3.85	2.37	....	1.43	....
Toronto	9.98	8.53	8.25	4.18	2.50	1.54	....	....	1.33		....	....
Blue Hill	8.61	8.89	6.16	4.57	3.66	1.60	0.65	....	1.62	1.54	....	0.61
Washington	9.51	9.53	7.41	4.80	3.82	2.40	1.80	3.73	2.28	1.20	....	1.13
Manila	10.63	11.64	6.42	3.90	4.64	2.32	1.49	3.14	1.82		....	....

Clearly, too, clouds cannot form at a lower level, the air there being unsaturated,—even if drifted in they would evaporate. Further, ordinary thermal convection usually does not extend to much higher altitudes, because the cooling of the rising mass through expansion and evaporation (the outer portions, at least, of the cloud evaporate) quickly brings it to or below the temperature of the surrounding air at the same level, except in the case of the largest cumuli, in which the amount of evaporation is very small in comparison to the total condensation. Hence foul weather cumuli and the lower cyclone clouds mark a second level of maximum cloudiness, commonly 1 to 2 kilometers above the surface.

Cumulus Level (b), *Fair Weather Type.*—During fair, calm, summer weather vertical convection is very strong but, as the relative humidity is low, the resulting clouds are of the alto-cumulus type. Hence the alto-cumulus, 3.5 to 4 kilometers above the surface, marks a secondary or fair weather cumulus level of maximum cloudiness.

Cirro-stratus Level.—Since the different types of cirrus formed in the region of a cyclone (the cirro-stratus being, perhaps, the most frequent) are spread far in advance of the storm itself by the swift upper winds, it follows that they also mark a level of maximum cloud frequency.

Cirrus Level.—During fair weather thin cirri often occur, as already explained, at or near the top of the troposphere, due, probably, to that marked cooling of the upper atmosphere characteristic of "highs," and as these are the highest of all clouds, it is obvious that they denote

a final level of maximum cloudiness, one whose average elevation in middle latitudes is about 10 kilometers.

Regions of Minimum Cloudiness.—Between the levels of maximum cloudiness there obviously must be regions of minimum condensation. These are:

Scud Region.—Since one level of maximum cloud formation, including fog, is at the surface of the earth and the next at an elevation of approximately 1.5 kilometers, the average base height of the cumulus, it follows that the intervening region is one of minimum cloudiness, the absolute minimum being just above the highest fog. The name "scud region" might be appropriate to this space, since "scud" is, perhaps, the only cloud that occurs in it.

Intercumulus Region.—The intercumulus region of minimum condensation lies, as the name suggests, between the cumulus and alto-cumulus levels of maximum cloudiness. Its elevation is, roughly, 2.5 to 3.5 kilometers.

Alto-stratus Region.—As the alto-cumulus and the cirro-stratus mark successive levels of maximum cloudiness at the heights of about 4 and 8 kilometers, respectively, it follows, as above that the region between at the heights of 4.5 to 6 kilometers especially the higher alto-stratus region, must be one of minimum cloudiness. And this it is, because (*a*) it is above the level of diurnal convection and therefore of most cumulus clouds; (*b*) the clouds of intermediate level formed in cyclonic areas are not blown forward so rapidly nor, therefore, over such wide areas as are the cirri; and (*c*) the atmosphere at this level in anticyclones is nearly always dry, apparently dynamically warmed, and therefore non-cloud-forming.

Intercirrus Region.—Since the cirrus region furnishes two successive levels of maximum cloudiness, a foul (cyclonic) and a fair weather type, whose elevations are about 8 and 10 kilometers, respectively, it follows that an intercirrus region of minimum cloudiness must lie between them at an elevation of, say, 8.5 to 9.5 kilometers.

Isothermal Region.—Obviously water vapor is not carried in any considerable amount beyond the limit of appreciable vertical convection. Hence, there being but little water vapor present, clouds cannot form in the stratosphere; that is, beyond an elevation of about 11 kilometers in middle latitudes.

There are, then, five principal levels of maximum cloudiness:

1. Fog level, surface of the earth or water.
2. Cumulus level, height above surface about 1.5 kilometers.
3. Alto-cumulus level, height above surface about 4 kilometers.

4. Cirro-stratus level, height above surface about 8 kilometers.
5. Cirrus level, height above surface about 10 kilometers.

There also are five regions of minimum condensation:

1. Scud region, 100 to 300 meters elevation.
2. Intercumulus region, 2.5 to 3.5 kilometers elevation, roughly.
3. Alto-stratus region, 4.5 to 6 kilometers elevation, roughly.
4. Intercirrus region, 8.5 to 9.5 kilometers elevation, roughly.
5. Isothermal region, beyond 11 kilometers elevation.

Cloud Depth or Thickness.—It is known that the thickness of clouds varies from the 8 or 10 kilometers of the most towering cumulus, usually associated with a violent hailstorm, down to that of a vanishingly thin cirrus. Systematic measurements of cloud thickness, however, have not been numerous. The best, perhaps, were made at Potsdam and are given in the following table copied from Hann's Lehrbuch der Meteorologie.

Cloud Thickness.

	Cloud	A.-St.	A.-Cu.	St.-Cu.	Nb.	Cu.-Nb.	Cu.	Fr.-Cu.
Depth in meters	Average	510	194	353	(590)	2070	669	214
	Maximum	1310	370	1265	1240	>4600	2230	430
	Minimum	105	50	50	160	340	90	70
Number of observations		6	18	18	16	21	22	26

Cloud Velocities.—The velocity of a cloud is the velocity of the air in which it floats, except in the case of a stationary type—crest cloud, banner cloud, et cetera—or a billow cloud. With these exceptions, it therefore is approximately the gradient velocity at the cloud level, which varies with altitude, latitude, temperature, and pressure distribution.

Average values, observed at certain places, are given in the following table also copied from Hann's Lehrbuch der Meteorologie:

Average Wind Velocity in Meters per Second.

	Ci.	Ci.-St.	Ci.-Cu.	A.-St.	A.-Cu.	St.-Cu.	Nb.	Cu.-Nb. Top	Cu. Top	Cu. Base	Fr.-Cu.	St.
1. SUMMER, CHIEFLY APRIL TO SEPTEMBER.												
Bossekop, 70° N	18	18	11	13	11	5	6	..	7	7	..	7
Upsala, 60° N	20	(39)	17	5	12	7	7	..	7	6	8	..
Potsdam, 52½° N	22	24	13	11	10	9	11	9	8	6	7	7
Trappes, 49° N	23	23	23	15	13	9	10	14	10	9	8	10
Blue Hill, 42° N	30	30	18	25	13	10	14	22	13	9	..	6
Washington, 39° N	30	27	23	18	16	10	8	15	7	..	..	6
Manila, 14½° N	13	16	3	..	11	4	..	..	..	..	..	..
Batavia (year), 6° S ..	12	19	3	..	6	6	..	..	4	..	..	1
2. WINTER, CHIEFLY OCTOBER TO MARCH.												
Upsala	23	13	18	..	13	12	6	18	12	..	12	..
Potsdam	28	20	24	16	16	12	13	28	10	(14)	12	10
Trappes	23	19	27	18	14	11	16	..	12	12	11	10
Blue Hill	37	41	36	25	24	13	13	..	..	15	..	10
Washington	35	30	33	21	21	15	12	21	11	..	..	10
Manila	13	16	3	19	4	8	6	..	..	..	..	..

CHAPTER VII.

ATMOSPHERIC OPTICS.

MANY curious and beautiful phenomena, of which the mirage, rainbow, halo, corona and sky colors are some of the more conspicuous, are due to the optical properties of the atmosphere and foreign substances suspended in or falling through it. Only a few of these, however, will be discussed here and they but briefly. Their practical importance generally is relatively small and a full discussion of almost any one requires a considerable knowledge of mathematics and physics.

MIRAGE.

It is a very common thing in flat desert regions, and during the warmer hours of the day, to see below distant objects and somewhat separated from them their apparent images, as though reflected from a horizontal mirror below the level of the observer—hence the name *inferior mirage* given to this phenomenon. It closely simulates, even to the quivering of the images, the reflection by a quiet body of water of objects on the distant shore—the "water" of course is the image of the distant low sky—and therefore frequently leads to the false assumption that a lake or bay is close by. This type of mirage is very common on the west coast of Great Salt Lake. Indeed on approaching this lake from the west one can often see the railway over which he has just passed apparently disappearing beneath a shimmering surface. It is also common over smooth paved streets provided one's eyes are just above the street level. An under-grade crossing in a level town, for instance, offers an excellent opportunity almost any warm day of seeing well defined small images that are apt to arouse one's surprise at the careless way his fellow citizens wade through pools of water!

Since the inferior mirage occurs only over approximately level places and there only when they are so strongly heated that for a short distance the density of the atmosphere increases with elevation, it follows that it is simply a refraction phenomenon. The velocity of light through the air increases, as is well known, with decrease of atmospheric density. Hence when a level surface is strongly heated the maximum density of the air and minimum velocity of light occurs at

a slight elevation. Clearly then, under such circumstances, rays from a given point on a low-lying distant object may reach the eye of the observer (when properly situated) along two very different paths, one slightly convex upward by which the object is seen substantially in its normal position, and another slightly convex downward by which the object is seen as reflected in the surface of a lake.

When the temperature of the air happens to increase very rapidly with elevation, as it sometimes does, for a short distance, the rays of light reaching an observer located in this air from a distant object are often appreciably curved downward. As a result of this, objects normally beyond the horizon "loom," or come into view; closer ones "tower," that is, assume unwonted heights, and occasionally also present upside-down images, the well knwon *superior mirage,* as though reflected from an overhead horizontal mirror. Indeed both the superior and inferior mirages are often, perhaps usually, "explained" as due to total reflection in the atmosphere. It is quite certain, however, that they are owing to refraction and not reflection, as simply and elegantly explained in Professor Hastings' book, Light (N. Y., 1902).

The practical importance of mirage phenomena, of which there are many, lies in their interference with angular measurements, and estimations of distance and direction. Indeed the uncertainty attached to the positions of objects seen in (or perhaps hidden by) a mirage is sometimes so great as to render observations of them worse than useless. An interesting case of this kind is noted in the report of the battle of April 11, 1917, between the English and the Turks in Mesopotamia, in which General Maude, the British commander, says: "The fighting had to be temporarily suspended owing to a mirage."

RAINBOWS.

It may seem entirely superfluous to describe so common a phenomenon as the rainbow, but rainbows differ among themselves as one tree from another, and besides some of their most interesting features usually are not even mentioned, and naturally so, for the explanations generally given of the rainbow, because of efforts to be simple, may well be said to explain beautifully that which does not occur, and to leave unexplained that which does. The ordinary rainbow, seen on a sheet of water drops—rain or spray—is a group of circular or nearly circular arcs of colors (usually said to be 7, but anyone who tries to count them is not likely to find so many), whose common center is on the line extended connecting the observer's eye with the source of light.

A very great number of rainbows are theoretically possible, though only 3 (not counting the supernumerary bows) certainly have been seen on sheets of rain. The most brilliant bow, known as the *primary*, with red outer border of about 42° radius and blue to violet inner border, appears opposite the sun (or other bright light); the next brightest, the *secondary* bow, is on the same side of the observer, that is opposite from the sun, but the order of its colors is reversed and its radius—about 50° to the red—is larger; the third, or *tertiary* bow, which has about the same radius as that of the primary and colors of the same order, lies between the observer and the sun, but is so faint that it is rarely seen in nature. The common center of the primary and secondary bows is angularly as far below the observer as the source (sun generally) is above it. Usually, therefore, less than a semi-circle of these arcs is visible, and never more except from an eminence. A record of close observations of rainbows soon shows that not even the colors are always the same. Neither is the band of any color of constant angular width, nor the total breadth of the several colors at all uniform. Similarly the purity and brightness of the different colors are subject to large variations. The greatest contrast perhaps is between the sharply defined brilliant rainbow of the retreating thunderstorm, and that illy-defined, faintly tinged bow that sometimes appears in a mist.

All these differences depend essentially upon the size of the drops, and therefore inequalities often exist between even the several portions, especially top and bottom, of the same bow, or develop as the rain progresses.

Rather narrow bands of color, essentially red, or red and green, often appear parallel to both the primary and secondary bows, along the inner side of the first and outer of the second. These are known as *supernumerary* bows. These also differ greatly in purity and color, number visible, width, etc., not only between individual bows but also between the several parts of the same bow. No such colored arcs, however, occur between the principal bows, primary and secondary. Indeed, on the contrary the general illumination here is perceptibly at a minimum.

No attempt will be made here to explain the rainbow. Any simple explanation, as already stated, is misleading. An accurate and adequate explanation of all rainbow phenomena is tedious and in some parts exceptionally difficult. It will be interesting, however, to consider a few of the more popular questions about the rainbow, as follows:—

(*a*) "What is the rainbow's distance?" In the sense of its proximate origin, the drops that produce it, it is near by or far away according to their respective distances, and thus extends from the closest to the farthest illuminated drops along the elements of the rainbow cone. Indeed the rainbow may be regarded as consisting of coäxial, hollow conical beams of light of different colors seen edgewise from the vertex, and thus having great depth or extent in the line of sight.

(*b*) "Why is the rainbow so frequently seen during summer and so seldom during winter?" Its formation requires the co-existence of rain and sunshine, a condition that often occurs during local convectional showers, but rarely during a general cyclonic storm, and as the former are characteristic of summer and the latter of winter it follows that the occurrence of the rainbow correspondingly varies with the seasons.

(*c*) "Why are rainbows so rarely seen at noon?" As above explained the center of the rainbow's circle is angularly as far below the level of the observer as the sun is above it, hence no portion of the bow can be seen (except from an elevation) when its angular radius is less than the elevation of the sun above the horizon. Now during summer, the rainbow season, the elevation of the sun at noon is nearly everywhere greater than 42°, the angular radius of the primary bow, or even 51°, the radius of the secondary bow. A rainbow at noon, therefore, is, except for very high latitudes, an impossible summer phenomenon, and, of course, a rare winter one, for reasons given above, even where possible.

(*d*) "Do two people ever see the same rainbow?" Theory teaches and ordinary experience shows, that as the observer remains stationary or moves, so also, other things being equal, does his rainbow. If then, two observers initially close together should move in opposite directions each would find his rainbow responding in the same sense as his shadow, and presently the positions and therefore the identity of the two bows would become unquestionably different, from which it follows that as the eyes of two observers must always be separated by a greater or less distance their bows must also be correspondingly separated and different,—different in the sense that they have different positions and are produced by different drops. In short, since the rainbow is a special distribution of colors (produced in a particular way) with reference to a definite point,—the eye of the observer,—and as no single distribution (other than uniform and infinite) can be the same for two separate points, it follows that two observers do not and can not see the same rainbow.

(*e*) "Can one see the same rainbow by reflection that he sees directly?" An object seen by reflection in a plane surface is seen by the same rays that, but for the mirror, would have focused to a point on a line normal to it from the eye, and as far back of it as the eye is in front. But, as just explained, the bows appropriate to two different points are produced by different drops, hence a bow seen by reflection is not the same as the one seen directly.

HALOS.

As is well known, cirrus clouds and others formed at temperatures considerably below 0° C. usually consist of tabular or columnar par-

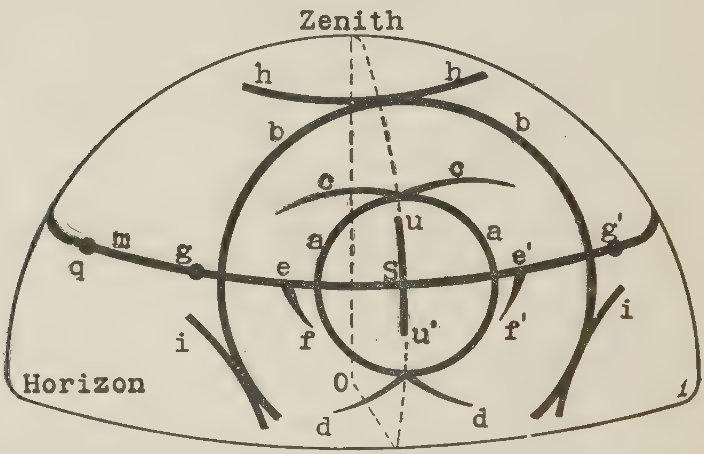

FIGURE 62. Perspective view of the sky, showing the sun (S); ordinary halo of 22° (a); great halo of 46° (b); upper tangent arc of the halo of 22° (c); lower tangent arc of the halo of 22° (d); ordinary parhelia of 22° (e, e′); Lowitz arcs (f, f′); parhelia of 46° (g, g′); circumzenithal arc (h); infralateral tangent arcs of the halo of 46° (i); the parhelic circle (m); a paranthelion of 90° (q); plane of the horizon; the observer (o).

ticles of ice, always hexagonal in type but varying greatly in detail. Light from the sun or moon obviously takes many paths through the clearest of such snow crystals, and produces in each case a corresponding and peculiar optical phenomenon. Other similar phenomena are produced by the reflection of light from the faces of these crystals. The former, or those due to refraction, are nearly all colored, the latter invariably white. All such phenomena, however, either pro-

duced by refraction or by reflection, are known as halos, the more common of which, as also one or two of the rarer forms, are illustrated in Figure 62, in which O is the position of the observer and S the position of the sun.

Referring to Figure 62, the circle *aa* around the sun and about 22° from it is the most common of the halos. Its colors are substantially those of the rainbow, red on the side next the sun and blue on the outer side. Within this halo the sky is less luminous than beyond it. The circle *bb* is known as the halo of 46°, that being its distance from the sun. It, too, is colored substantially as the halo of 22°, but is less brilliant and far less frequently seen. *hh* represents what is known as the circumzenithal arc. It is parallel to the horizon, generally a little more than 46° from the sun at its nearest point, and has its center at the observer's zenith. It never forms a complete circle, and indeed rarely more than one-fourth to one-third of a circumference. It appears only when the sun's altitude is less than 31°, and generally can be seen for only a few minutes. It is, however, the brightest perhaps of all halos and frequently is erroneously reported as a brilliant but unusual rainbow. A faint arc, known as Kern's arc, is occasionally seen diametrically opposite the circumzenithal arc, as though a portion of the same circumference. *cc* and *dd* are known as the upper and lower tangent arcs, respectively, of the halo of 22°. They undergo great changes in extent and position with the altitude of the sun. At certain altitudes they merge together into what is sometimes mistaken for a circumscribing ellipse. *i* and *i* are known as the infralateral tangent arcs of the halo of 46°. *ee'* and *gg'* are the parhelia of 22° and 46° respectively. When the sun is on the horizon they are on the circumference of the circular halos of 22° and 46°. With elevation of the sun, however, they separate from these circles farther and farther. *f* and *f'* are known as the arcs of Lowitz, rarely seen distinctly and rather difficult and tedious of explanation. *qgSg'*, etc., is the parhelic circle. It is produced by reflection of light from the faces of snow crystals, is colorless, and passes, as its name implies, through the parhelia or mock suns and is parallel to the horizon with the zenith of the observer for its center. The vertical line *uu'* is known as a light pillar, and is also produced by reflection and therefore white. These are only some of the possible and occasionally seen halos, but include the more common ones. Together they form an interesting study, but can not be adequately discussed without a considerable knowledge of optics and a rather free use of mathematics.

CORONAS.

Coronas consist of one or more of rainbow-colored rings, usually of only a few degrees radius, concentric about the sun, moon or other bright object when covered by a thin cloud veil. They differ from halos in having smaller (except in rare cases) and variable radii, and in having the reverse order of colors; that is, blue nearest the sun and red farthest away. They owe their origin, not to refraction by ice crystals, but to diffraction by small water droplets—the larger the droplets the smaller the radius of the corona. From the angular size of a corona it is easy to compute the approximate diameter of the particles that produce it, but the theory upon which this computation is based, like all the other phenomena of coronas, rainbows, halos, etc., requires special mathematical discussion that generally may be found only in advanced work on optics. Those therefore who wish to follow up any of the fascinating topics of meteorological optics must have recourse to special treatises, most of which are difficult to read, but which reward the necessary labor to master them.

CHAPTER VIII.

GENERAL CIRCULATION OF THE ATMOSPHERE.

INTRODUCTION.

ATMOSPHERIC circulation, whether manifesting itself in a monsoon or in only a gentle lake breeze, is a gravitational phenomenon induced and maintained by temperature differences. This can be well illustrated by the flow of water between two adjacent tanks when connected by an upper and a lower pipe and kept at different temperatures.

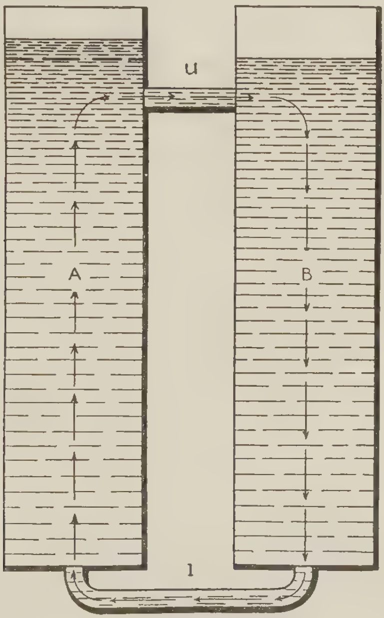

FIGURE 63. Circulation between warm and cold tanks.

Let the two tanks, A and B, Figure 63, be filled to the same level slightly above the upper pipe *u,* and let them have the same temperature. Under these conditions there will be no flow of water from

either tank to the other. Now let the pipes be closed and let the water in tank A be equally warmed throughout. It will expand, provided its original temperature was not below 4° C., and the amount of water above each level in A, at and below the initial surface, be increased in proportion to its distance from the bottom. Hence the pressure due to gravity is everywhere through the original volume correspondingly increased—the maximum increase being at the level of the initial surface. If the lower pipe *l* be now opened, there still will be no flow of water from either tank to the other. But if the upper pipe be opened, water will flow from A to B, and in so doing will decrease the pressure on all parts of A and increase it on all parts of B. If *l* is also open, water will flow from B to A. If both pipes are left open and the water in A kept constantly warmer than the water in B, there will be continuous circulation of the water from A to B through the upper pipe and from B to A through the lower. Obviously the same results could be obtained by applying a cooling process to B instead of a warming one to A. That is, since the circulation in question is a gravitational phenomenon induced by a temperature difference between the water in the two tanks, it clearly is immaterial how this temperature difference is established, whether by heating the one tank or by cooling the other. Similarly in the case of the atmosphere. If two adjacent columns of air, or the masses of air over two adjoining regions, whether large or small, are kept at different temperatures, there will exist, through the action of gravity, a continuous overflow from the warmer to the colder, and an underflow from the colder to the warmer. Neither does it make any difference in this case how the inequality of temperature is established and maintained, whether by heating the one section or by cooling the other.

Clearly then, as the tropical belt of the earth generally is the warmest, and the higher latitude regions on either side the coldest, there must be, and are, two branches of the general or planetary circulation of the atmosphere, covering respectively the northern and the southern hemispheres. If then the earth were smooth (had no mountains or other elevations), if its temperature varied uniformly and equally from equator to either pole, and—most important of all—if the earth were non-rotating, the general circulation of the atmosphere would consist of a steady flow of the upper air towards the poles and of the lower air towards the equator. But this ideal simplicity does not exist in any particular, and consequently the actual circulation of the atmosphere differs very widely from the simple type indicated.

WINDS IN GENERAL.

Effect of Earth Rotation.—From the fact that the linear velocity of the surface of the earth decreases from about 1040 miles per hour at the equator to nothing at the poles it follows that the equator-poleward and return circulation must be thrown very much askew, the poleward-flowing portion trending eastward and the return westward. The result of this is that in middle latitudes the prevailing winds are from westerly points, and in equatorial regions from easterly points. When the equatorial air, with an east to west velocity component, moves toward a polar region, as it does under the influence of the great temperature contrasts between equatorial and polar regions, it might seem, as many have assumed, that the air would tend to retain its original linear velocity around the axis of the earth. But this, too, is another case where in reality matters are not so simple as at first they may seem.

Although winds generally between latitudes 30° N. and S., especially over the oceans, are from easterly points, their westward velocity with reference to the surface of the earth is only a small fraction of the actual eastward velocity of the surface. Hence, though blowing from the east, the air in reality is rotating about the earth's axis in the opposite direction, or from west to east, though not quite so rapidly here as the earth itself. As this air moves to higher latitudes it comes closer and closer to the axis about which it is rotating—the axis of the earth. Similarly air that moves to lower latitudes gets further from this axis. Hence what it known as the law of the conservation of areas applies, except as modified by friction, etc., to the interzonal circulation. Or, in the terms of the physicist, the angular momentum, $mr\omega$ (in which m is the mass, r the distance from the axis of the earth, and ω the angular velocity of the earth's rotation), tends to remain constant, and not the linear momentum, mv (in which v is the west to east component of the linear velocity).

This subject, like so many others in meteorology, is not simple, but it is vital to an understanding of the circulation of the atmosphere. Those especially interested might look up its discussion in works on the physics of the air.

Automatic Adjustment of Winds in Direction and Velocity.—In discussing the more extensive winds it is convenient to consider the earth as stationary and the air as moving over it without friction under the influence of three distinct horizontal forces: (1) The deflective force, due to the earth's rotation; (2) the horizontal component of the

centrifugal force, due to the curvature of the path, and (3) the horizontal or gradient pressure, due to gravity. The first two are at right angles to the course of the wind and therefore help to control its direction, but do not alter its speed. The latter, however—that is, the gradient pressure—affects both the direction and the speed. Furthermore, as the velocity depends upon the horizontal pressure alone, and as the other forces depend in turn upon the velocity, and are zero when it is zero, it follows that of the three forces only the gradient pressure is independently variable.

Consider, then, the result of applying a horizontal pressure p of constant magnitude and constant geographic direction to a small mass m of air, free, as above assumed, from friction: Let m, Figure 64, be the mass in question initially at rest with reference to the surface

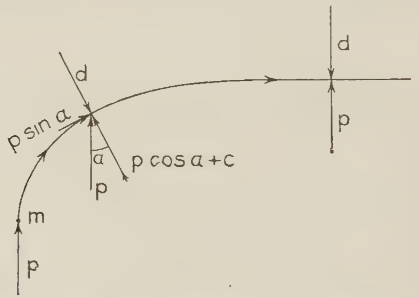

FIGURE 64. Deflection and path of winds in frictionless flow under a force of constant magnitude and constant geographic direction.

of the earth, and let it be acted on by the force p, exactly poleward, say. Immediately the mass moves, under the applied pressure p, the deflective force d becomes operative, thus curving the path (to the right in the northern hemisphere, to the left in the southern) and introducing the centrifugal force c. So long, however, as the angle between the path and the force p is less than 90 degrees there will still be a component of the latter in the line of motion; accordingly the speed of m will continue to increase, and therefore also the deflective force d. If this angle should exceed 90 degrees, the force p would have a component opposite to the direction of motion, which consequently would be slowed up and d thereby correspondingly decreased. In the end, therefore, a poleward force along the meridians on an object free to move gives it an exactly west to east velocity of such magnitude that, except in very high latitudes, the resulting deflective force is nearly equal to the horizontal pressure—the horizontal component of

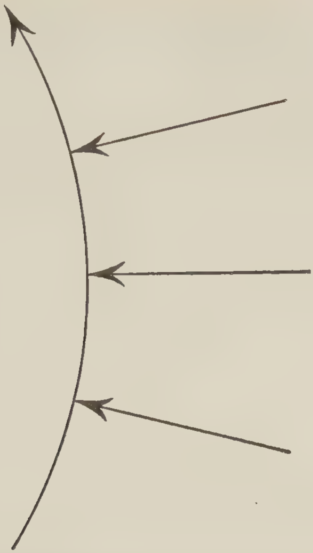

FIGURE 65. Path of winds in frictionless flow under a converging force.

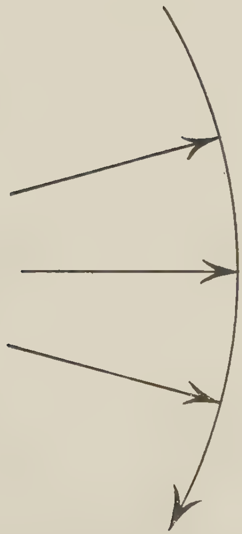

FIGURE 66. Path of winds in frictionless flow under a diverging force.

the centrifugal force being then comparatively small, except near the poles. Whenever the direction of the gradient force, whether poleward, as above assumed, or any other, the final motion is normal thereto.

A change in the magnitude but not in the direction of p, above, would only shift the latitude of the path and change the velocity so as to be nearly proportional to p.

If the horizontal pressure is not everywhere in the same direction, but converges, as in Figure 65, or diverges, as in Figure 66, the path, in adjusting itself normally to the directions of this pressure, obviously curves, as in cyclonic and anticyclonic regions, respectively.

In all cases, then, the wind automatically follows approximately the isobar of its position, with substantially the gradient velocity.

General Relations of Wind to Elevation.—Knowledge of the directions and velocities of the winds of the earth is still fragmentary and incomplete. Over large areas even the surface winds are unknown, and over regions best studied these alone are well known. The continuous records obtained at mountain stations have given much information in regard to air movements, but stations of this nature are comparatively few and, besides, their data, however valuable, are always affected to an unknown extent by local topography. Cloud observations have also given a large amount of valuable information, but it, too, is only fragmentary. At best a cloud observation seldom gives more than the direction and velocity of the air at one level, nor does such an observation ever apply to the stratosphere, since this region is never visited by clouds. In many respects kites and sounding balloons have furnished the most valuable data in regard to the movements of the upper air and their causes, but, unfortunately, aerological investigations of this nature, with relatively few exceptions, have been restricted to the northern hemisphere, and even there mainly to the summer season. Nevertheless, by combining the data gathered from these various sources a number of tentative conclusions, subject, of course, to modification, have already been reached in regard to the winds of different parts of the world from the surface up to great elevations. Some of the more important of these conclusions are:

1. That there is no continuous and rapid overflow of the atmosphere at all longitudes from the equatorial to the polar regions. At an elevation of 10 kilometers, for instance, the wind of middle northern latitudes seems to have southerly components about as often as northerly.

From this it follows that the equator-polar circulation is irregular and probably complex even at the higher altitudes.

> 2. That the equatorial winds are not always and at all levels from the east; that, on the contrary, west winds occur (how regularly is uncertain) at elevations of about 18 to 20 kilometers, with east winds again prevailing (certainly at times) at still greater elevations.

The cause of this layer of equatorial west wind has never been explained. Indeed, it may be only a local and temporary phenomenon.

> 3. That layers of air in which the temperature increases with increase of elevation, and others in which the temperature is constant, exist at different levels, especially through the first two or three kilometers. This stratified condition of the lower atmosphere appears to be universal. It is found even over tropical oceans, and is exceedingly well developed over the ice plateau of Antarctica.

Each layer usually shows such different humidity and such different wind velocity from those of the adjacent layers as to indicate a distinct origin, which it well may have. A rising convection current on reaching its equilibrium level flows away substantially at that particular elevation, and obviously retains its own humidity (provided condensation has not taken place), dust content, and other peculiarities. Its viscosity is not the same as that of the adjacent air, because its humidity or temperature, or both, are different. Hence, as shown by billow clouds, any such layer with a distinctly independent velocity tends to retain its integrity and to glide over another from which it differs physically without rapid intermingling. And there are still other obvious causes of temperature and humidity irregularities and consequent stratification of the atmosphere, such as reflection from, and evaporation of, clouds, surface cooling, and air drainage. Clearly, then, one should expect to find in the lower atmosphere substantially the kind and amount of temperature inversions and other irregularities that it actually shows.

> 4. That the upper winds are exceedingly variable along the edges of the high-pressure belts, and that marked disturbances occur in the antitrades.
>
> 5. That the north-poleward pressure gradient in the upper atmosphere becomes very small long before the arctic circle is reached—in fact, between 50° and 60° N.

6. That in high northern latitudes, where the poleward pressure gradient of the upper atmosphere is small, the westerly winds are not constant.

Season of Greatest Winds.—From the above discussion it is obvious that the general wind will be swiftest whenever the temperature contrast between the air of higher and lower latitudes is greatest. But the temperature of the atmosphere in low latitudes does not change through the year nearly so much as does that of higher latitudes. Hence, the maximum horizontal temperature gradient, and therefore the greatest pressure gradient and strongest winds, must occur during winter.

Latitude of Greatest Winds.—The latitude of strongest winds clearly is that at which the horizontal pressure gradient is greatest. In the northern hemisphere, according to Figure 43, this occurs in the summer at about latitude 45°. It is obvious, however, since the pressure gradient depends in general upon the latitude rate of temperature change, that the belt of maximum winds must shift more or less from season to season—poleward with the coming of summer, equatorward with the onset of winter.

Hours of Greatest and Least Winds.—On land, but not appreciably at sea, the velocity of the surface wind has a well-defined daily period. Over comparatively level regions it is least, on the average, about sun-up and greatest from 1 to 2 P. M., with a larger change on clear days than on cloudy. It is also most pronounced in summer, when it reaches an average altitude of about 100 meters, and least in winter, when its depth is only about 40 meters.

The physical explanation of this phenomenon was given long ago by Espy. During the night, when there is no vertical convection, surface friction holds the lower air comparatively quiet, while the upper air glides over the lower with but little restraint. During the day, however, and especially during clear, summer days, vertical convection causes the surface layers of air and those directly above to become thoroughly mixed and therefore to have a more or less common velocity, which, obviously, is greater than the undisturbed or night surface velocity, and less than the undisturbed upper layers before their mixture with the lower.

Daily changes of wind velocity also occur on mountain tops, where the maximum is at night and the minimum by day, or just the reverse of the velocity changes that occur near the surface over plains. Three factors, possibly more, combine to produce this result: (*a*) Contrac-

tion of the lower air by night, thus bringing air of slightly higher levels, possibly 15 meters (50 feet) or so, and therefore of somewhat greater velocity down to the mountain top. (*b*) The presence by day and absence by night of surface disturbance, due to convection, in the air flowing over the mountain. (*c*) Overflow from the region of maximum expansion to the region of maximum compression. Since the greatest expansion usually occurs at 3 to 4 P. M. and the greatest compression at 5 to 6 A. M., it follows that the overflow will be from west to east, or with the prevailing winds, through the night, and from east to west, or against them, during most of the day; that is, from sun-up to 3 or 4 P. M.

Daily Direction of the Wind.—The average direction of the wind changes slightly during the day, both over plains and on mountain tops, the tendency being for it always to follow the sun, or, rather, the most heated section of the earth. That is, the wind tends to be east during the forenoon, south (in the northern hemisphere) during the early afternoon, and west during the late afternoon and early evening. This does not mean that at each instant the wind really blows directly from the then warmest region, but that the actual changes through the day in the average hourly wind directions can be accounted for by a velocity component away from that region. The whole sequence results from the thermal expansion of the atmosphere (progressive from east to west), which causes an increase of pressure and consequently an outward flow at all levels above the surface. The area covered is so vast that the time involved, only a few hours, is insufficient for the completion of the convention circuit, so that even the surface winds are *away* from the most heated regions, as stated, and not toward them, as in sea and land breezes, for instance. The compensating or return current occurs at night, when the component, outside the tropics at least, is from the higher latitudes. In reality the entire phenomenon is only a diurnal surge, a flux and reflux, of the atmosphere due to diurnal heating and cooling.

Normal State of the Atmosphere.—From the above explanations of the causes of general winds, it appears that the normal state of the atmosphere is one of considerable velocity with reference to the surface of the earth. In middle latitudes, at least, this velocity is from west to east more or less along parallels of latitude and so great as nearly to balance the latitudinal pressure gradient due to the zonal distribution of insolation. Calms, therefore, in this region must be regarded as disturbances of the atmosphere, and indeed often are comparatively shallow, with normal winds above.

Equatorial East to West Winds.—East to west winds are quite as general and constant in equatorial regions as are west to east winds in middle latitudes. Along its borders, roughly 30° N. and 30° S., this equatorial belt of east to west winds is very shallow. Toward the equator its thickness increases, as a rule, until it reaches at least the limit of vertical convection. There are, however, great irregularities in these winds, just as in those of higher latitudes on either side of it. But the general conditions are as stated.

Probable Interzonal Circulation of the Stratosphere.—The primary circulation just explained involves all the atmosphere from the surface of the earth up to at least the highest cloud levels, but there is reason to believe that it does not extend to the greatest altitudes. Indeed, it appears probable that far above the uppermost clouds there may be another primary or fundamental circulation in reverse direction to that of the lower. This inference is based on the fact that the stratosphere is so much warmer in high than in low latitudes that seemingly there must be an overflow of air from the former to the latter and a corresponding return; that is, a primary circulation in the stratosphere in which the upper branch is from the polar (in this case warmer) toward the equatorial (in this case colder) regions and the under from the equatorial toward the polar regions, with, of course, longitudinal components in each due to the earth's rotation. In a sense the upper circulation, if it exists as inferred, is the mirror image of the lower, though more regular.

MONSOONS.

Summer monsoons and winter monsoons, for convenience discussed under the same head, bear the same relation to summer and winter that sea breezes and land breezes bear to day and night. It is the temperature contrast between land and water that establishes the circulation that manifests itself on the surface as a sea or land breeze in the one case and as a seasonal or monsoon wind in the other. The direction of the surface wind in either case is always from the cooler toward the warmer of the adjacent regions, from the ocean toward the land by day as a sea breeze and during the warmer season as a summer monsoon; from the land toward the ocean by night as a land breeze and during the colder season as a winter monsoon. Hence monsoons may be regarded as sea and land breezes of seasonal duration, and might very well be classed with the latter under some common appropriate caption. However, because of the immense areas

involved, it cannot be said of them, as of sea and land breezes, that they are caused by mere local temperature differences. Besides, the duration of a land or sea breeze is so brief that it covers only a narrow strip along the coast, as already explained, while the monsoon winds extend far from the coast, both inland and to sea and the directions of the former, since their paths are always short, are but little affected by the rotation of the earth, while the courses of the second are greatly modified by this important factor.

The prevailing directions of monsoon winds, except where distinctly modified by the general circulation, are given by the following table.

Direction of Monsoon Winds.

Hemisphere	Season	Land south	Land west	Land north	Land east
Northern	Summer	N. E.	S. E.	S. W.	N. W.
	Winter	S. W.	N. W.	N. E.	S. E.
Southern	Summer	N. W.	N. E.	S. E.	S. W.
	Winter	S. E.	S. W.	N. W.	N. E.

Since monsoons depend upon seasonal temperature contrasts between land and water, it is obvious that winds of this class must be most pronounced where such contrasts are greatest—that is, in temperate regions—and least developed where the temperature contrasts are smallest—that is, in equatorial and polar regions. It is even possible for secondary monsoons to develop, or for a monsoon to occur within a monsoon. This merely requires a favorably situated inland sea, such as the Caspian. In such cases monsoons or seasonal winds prevail between the inland sea and the surrounding land, and in turn between the continent as a whole and the adjacent oceans, just as and for the same reason that on a still greater scale, there is a constant circulation between the perpetually warm equatorial regions and those about the poles that are continually cold.

Another comparison between these several winds, the semi-daily (land and sea breeze), semi-annual (monsoon), and perpetual (interzonal), that is interesting and instructive concerns their depth. As already stated, the land and sea breezes seldom reach greater depths than 100 to 500 meters; the winter monsoon of India has a depth, roughly, of 2000 meters, and the summer monsoon 5000 meters; while the general or interzonal circulation involves the whole of the troposphere with a depth of 10 to 12 kilometers, and probably also, though perhaps to a less vigorous degree, even the stratosphere.

If the term monsoon be extended, as it properly may, to include all winds whose prevailing directions and velocities undergo distinct alterations as a result of seasonal changes in temperature, it clearly

follows that this class of winds is well nigh universal. Nevertheless, it is generally thought of in connection with only those places where it is most strongly developed, and especially where the seasonal winds are more or less oppositely directed. Among these places are: India (Indian monsoons are the most pronounced of all and have been most fully studied), China, the Caspian Sea, Australia, and portions of Africa.

In the United States the chief monsoon effects are in the eastern portion, where the prevailing winds are northwest in winter and southwest in summer, and in Texas, where the prevailing winds are also northwest in winter but southeast in summer.

TRADE WINDS.

As previously stated, in equatorial ocean regions, or, roughly, over the oceans between latitudes 30° N. and 30° S., the winds usually have an east-to-west component. In the northern hemisphere they blow rather constantly from the northeast, becoming east-northeast and finally nearly east winds as the equator is approached. Similarly, in the southern hemisphere, starting from the southeast, they gradually back through east-southeast to nearly east. In each case they blow "trade"; that is, in a fixed or nearly fixed direction. It is because of this steadiness of direction and not because of any relation they may have to the paths of commerce that they are called trade winds. Along each border of this belt, or along both the northern and southern horse latitudes, calms are frequent, while such winds as do occur generally are light and variable in direction. Besides, the barometric pressure is high, humidity low, and sky clear. Hence it generally is inferred that throughout the horse latitudes the air is descending. This evidence, however, as applied to places other than the centers of maximum pressure is not quite conclusive—it only shows that the air is not ascending.

Another narrow belt of calms or light variable winds, known as the region of the doldrums, approximately follows the equator (more exactly the thermal equator), where the two systems of trade winds, the northern and the southern, come together. Here, however, the barometric pressure is low, humidity high, and skies often filled with cumulus and other clouds that give conclusive proof of strong ascending currents.

Trade winds in the sense here used—that is nearly constant winds blowing in a westwardly direction—do not occur on land except along coasts and over islands. Besides being well-nigh peculiar to the oceans,

they are even different from ocean to ocean, and also, since they tend to follow the thermal equator, somewhat different in latitude and intensity from season to season.

According to Shaw the average velocities of the Atlantic trade winds are as follows:

Trade-wind Velocities, Atlantic Ocean.

	Jan.	Feb.	Mar.	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year	
N. E. trade ...	11.9	13.0	13.5	13.4	12.3	11.4	10.3	8.3	9.6	7.4	9.8	11.6	10.5	miles/hour
S. E. trade ...	14.1	13.0	13.0	12.1	11.0	12.1	12.1	15.0	17.0	15.0	16.1	15.0	13.9	miles/hour

From this it appears that the trades are strongest during the winter when their counterpart, the system of westerly winds of higher latitudes, is strongest; and weakest during the summer when their counterpart is weakest. It also appears that the southeast trades, or those pertaining to the southern hemisphere, are about one-third stronger than the northeast trades, owing probably to the greater extent of the southern oceans and consequent less surface friction—the same reason, doubtless, that the westerly winds of the southern hemisphere are stronger, on the average, than the westerlies of the northern hemisphere.

The trade winds of the Pacific Ocean are weaker than those of the Atlantic and not so constant in direction. On the Indian Ocean the trades are confined to the southern hemisphere. North of the equator the winds of this ocean, being controlled by the adjacent continent, are distinctly of the monsoon type.

The seasonal shifting of the trade regions and belt of doldrums is shown by the following table, copied from Hann's Lehrbuch, 3d edition, p. 463:

Seasonal Latitude Limits of Trade Winds and Doldrums.

	March		September	
	Atlantic	Pacific	Atlantic	Pacific
N. E. trade	26°–3°N.	25°–5°N.	25°–11°N.	30°–10°N.
Doldrums	3°N.–Equator	5°–3°N.	11°–3°N.	10°–7°N.
S. E. trade	Equator–26°S.	3°N.–28°S.	3°N.–25°S.	7°N.–20°S.

ANTITRADE WINDS.

As the heated and expanded air of equatorial regions overflows to higher latitudes it necessarily is deflected by the rotation of the earth. That portion which goes north changes from an east wind near the equator to a southeast, south, southwest, and, finally, at about latitude 35° N., a more nearly west wind. Similarly, that portion which goes south becomes northeast, north, northwest, and, finally, at about latitude 30° S. a more nearly west wind.

At great altitudes, 10 to 15 kilometers, the east-to-west velocity near the equator is, roughly, 36 meters per second (80 miles per hour). Hence its west-to-east velocity around the axis of the earth is about 428 meters per second (957 miles per hour). As this air, assuming it to start from the equator and neglecting viscosity effects, moves to higher latitudes its west-to-east velocity must so increase, according to the law of the conservation of areas, that at about 16° N. or S. its angular velocity will be the same as that of the earth, and itself, therefore, be moving only poleward in the plane of the meridian. The exact latitude, however, at which the antitrades move directly poleward depends upon the position of the thermal equator and therefore varies with the seasons. Thus during August and September, when the center of the doldrums is, roughly, 8° N., the inflection of the northern antitrades occurs somewhere between latitudes 20° N. and 25° N. At other seasons, because the doldrums are then nearer the equator, the place of inflection is also less removed. Beyond the turning point, wherever that may be, these upper or antitrade winds become westerly, and, except as modified by local disturbances, tend, as previously explained, to reach, under the influence of the poleward pressure, a limiting or gradient velocity and to follow parallels of latitude. However, there are innumerable disturbances, mainly due to the distribution of land and water, that cause constant and abundant interzonal circulation which feeds and indefinitely maintains the antitrade wind portion of the general or planetary atmospheric circulation.

The height of the antitrades (depth of the trades) is greatest, at any given place, during summer and least during winter. It also decreases with latitude, becoming zero, on the average, at about 30° N. and S. Thus during winter their height over Cuba, 22° N., is about 3.5 kilometers; over Jamaica, 17° N., 6.5 kilometers; over Trinidad, 12° N., 8 kilometers, and over Hawaii, 19° 30′ N., about 3 kilometers. But whatever their height it is always the same as the depth of the trades of which they are but the overhead continuation. Indeed, the trade winds as they approach the equator ascend and gradually flow off poleward, thus producing in each hemisphere a great antitrade branch of the general circulation, which in turn becomes the westerlies of higher latitudes. These, in their turn, are confused by storms and other local disturbances, but after few or many vicissitudes, as circumstances may determine, ultimately return to a similar starting-point, only to begin another of their endless cyclic journeys through trades, antitrades, westerlies, and the innumerable secondary winds that such a course implies.

CHAPTER IX.

SECONDARY CIRCULATION OF THE ATMOSPHERE.

INTRODUCTION.

In middle latitudes the weather, which affects the well-being of animal and vegetable life from day to day, is largely the result of moving areas of low and high barometric pressure known respectively as cyclones and anticyclones. These are really large disturbances in the atmosphere and their influence covers a district from a few hundred to a thousand or more miles in diameter.

In the temperate zone these disturbances move eastward with an average velocity of from 22 to 37 miles an hour. The rate is somewhat greater in winter than in summer and the movement of cyclones slightly more rapid than that of anticyclones. Since the barometric pressure decreases toward the center of the cyclone there is a movement of the air at the surface of the earth more or less in the same direction. This inward flowing surface air does not move directly toward the center, however, but 60° or so to the right of it in the northern, and to the left of it in the southern hemisphere, owing to the rotation of the earth. With increase of elevation, this angle increases, until at an elevation of 1500 feet or even less, it is practically 90°. That is, at and beyond this level, the winds follow isobars, nearly, instead of blowing across them as one might suppose. As the lower inflowing air approaches the center of the cyclone there is a decided upward component, especially on the eastern side.

It is important that the extreme thinness of these areas be appreciated as compared with the horizontal distances over which their influence extends. This important fact must not be lost sight of in connection with the daily weather maps, such as Figure 67, p. 121. Here the cyclone centered at Philadelphia, influenced the surface winds over an area nearly 1,000 miles in diameter while all this air movement was really in the form of a relatively thin disc—the vertical dimensions being small as compared with the horizontal.

A reference to the weather map, Figure 67, p. 121, as well as to other daily weather charts, will show that the surface wind move-

ment under the influence of the "lows" is spirally inward toward the center, and, under the influence of the "highs," spirally outward. In the northern hemisphere the spiral movement in the "lows" is opposite to the movement of the hands of a watch laid face up, while the air flows out from the "highs" in a spiral motion with the hands of a watch. In the southern hemisphere the circulation of the air in these areas is in the opposite directions. There are comparatively few cyclones and practically no anticyclones within the tropics or even between 30° on either side of the equator. Tropical cyclones move westward while within 20° or so of the equator. At higher latitudes they usually recurve and turn eastward.

Two important general laws can be set down, therefore, in connection with cyclones and anticyclones. (1) In temperate latitudes cyclones and anticyclones move from the west toward the east. (2) Surface winds are controlled by difference in pressure, the surface air flowing spirally in *toward* the cyclone and spirally out *away* from the anticyclone.

An inspection of weather maps will show further that the temperature is higher to the east and south of cyclones than it is to the north and west of the center, because the temperature of the air is influenced by the temperature of the region over which it moves. Hence the third important law. (3) The direction of the surface winds largely controls the temperature at any place on the surface of the earth.

WEATHER.

A moving mass of air that has an upward component, whether flowing over uneven ground or moving upward near the center of low pressure areas, expands, and is cooled adiabatically at the rate of 1.6° F., for each 300 feet of ascent; descending air is warmed at the same rate. As the capacity of air for moisture depends almost exclusively on its temperature, it follows that ascending air is more apt to be cloudy and rainy than descending air. If a current of air with a temperature of 80° and a relative humidity of 75 per cent. is forced up 1990 feet, the temperature of condensation, allowing for increase of volume, will be reached and clouds will form. If the process is continued, rain may be expected, the amount of precipitation depending upon the quantity of air flowing in, the moisture content, and the extent of cooling.

In general, then, cloudy and rainy weather may be expected when air is ascending, while clear skies will be anticipated when the air is descending. With the relation between pressure and wind in mind,

it will be readily seen that lows are generally accompanied by clouds and rain, while large areas of high pressure will usually be accompanied by clear skies.

In the middle and eastern United States, the heaviest rains are found on the eastern side of the cyclones, because here the southerly winds are usually warm and moisture-laden. As they ascend in approaching the center of the low pressure area, adiabatic cooling takes place and the temperature of condensation is soon reached. If the movement of the cyclone is slow, large masses of moisture-laden air will flow toward it from the south and heavy rains may be expected at some distance southeast of its center. Sometimes the cyclone will be nearly stationary, when the rains may be extensive and floods result. Usually the eastward movement of the low will cause the area of rainfall to spread eastward and give generous rains over large districts. Occasionally a succession of energetic cyclonic areas will follow each other closely, each with excessive rains. This was the case in March, 1913, when the rivers in the northern Ohio watershed were from 10 to 16 feet higher than ever before recorded, owing to excessive downpours of rain.

COLD WAVES.

When a well defined and energetic cyclonic area moves eastward across the central Mississippi Valley and the Great Lakes, strong southerly winds to the south and east of the center will cause unseasonably high temperatures, especially in winter time. With the shift of wind to west and northwest, as the center of disturbance moves eastward, the temperature falls rapidly. When the approaching high is large and well defined, the northwest winds, often accompanied by snow squalls, are strong and the fall in temperature in 24 hours sometimes amounts to 40° or 50° or even more. These are the conditions which make up the well known winter cold wave of the United States. After the windy front of the anticyclone has passed and the center lies over a district, the nighttime temperatures will be very low, especially in the valleys, under the influence of radiation.

The conditions of high and low pressure follow each other with a fair degree of regularity and at an average velocity of 600 miles in 24 hours. About two anticyclones and two cyclones may be expected to pass over any point in central and northeastern United States each week, with the attendant shifts in wind direction, consequent changes in temperature, and varying weather conditions.

THUNDERSTORMS.

All the features of thunderstorms point to their dependence on a convectional overturning of the atmosphere. They most frequently occur in warm regions and are most common in spells of warm summer weather and in the afternoon at, or shortly after, the hour of the day when convectional movements are most active.

Thunderstorms will usually occur wherever there is a rapidly rising current of moisture-laden air. These conditions may obtain in a large current of air moving up the side of a mountain or in rapidly ascending currents of air in a comparatively level region under conditions of unstable equilibrium. The unstable condition may be produced by the overheating of surface air or the excessive cooling of air aloft. The overheating of surface air is brought about in hot summer weather when the air is comparatively quiet and the surface of the ground becomes greatly heated by uninterrupted insolation. The storms which occur under these conditions are classed as "heat" thunderstorms. They are sporadic in character and seldom continue for any great length of time. They may move slowly in any direction.

A second class of thunderstorms occurs principally in southerly winds to the eastward of easterly moving cyclonic areas. These are also generated by the unstable conditions produced by the overheating of a large mass of surface air or else in currents of air that are deflected upward by hills or shore lines. The formation of cumulus clouds is a common occurrence in this region, and these frequently grow into cumulo-nimbus or so-called thunderheads. It is thunderstorms of this type which frequently show the anvil-shaped clouds and which are so commonly accompanied by hail.

A third class of thunderstorms occurs in a V-shaped trough or depression extending southwesterly from a cyclonic area, where warm southerly winds are replaced by colder winds from the west. This is particularly a region of unstable equilibrium. It is thunderstorms of this class that are most likely to be followed by several days of cooler weather, due to the advancing cool wave behind the cyclone, while the cooling that accompanies the first and second class of storms is usually only temporary.

In the third class of storms, the wind squall which accompanies them is often strong enough to level trees and overturn frail buildings. These wind squalls are often spoken of as "tornadoes" but instead, they are straight-line thunderstorms.

A fourth class of thunderstorms should be included, although they are much less frequent. These occur in winter and sometimes in snow-

storms. Severe lightning damage sometimes occurs with these storms, evidently due to the fact that the region of greatest electric accumulation is but a few hundred feet above the surface of the ground.

Thunderstorms usually move toward the east in the temperate region and toward the west in the torrid zone. In middle latitudes the velocity of progression, except in the case of the slow-moving heat thunderstorm, is from 20 to 50 miles an hour and is somewhat greater than that of the cyclone which they may accompany.

TORNADOES AND WATERSPOUTS.

The tornado is the most diminutive and yet the most violent and destructive of all storms. It may be defined as a violent wind storm acompanied by hail, thunder and lightning, in which the air masses whirl with great velocity about a central core while the whole storm travels across the country in a narrow path at a considerable speed. When seen from a distance the tornado has the appearance of a dense cloud mass, usually in violent agitation and with one or more pendant funnel-shaped clouds, which may or may not reach the earth. Waterspouts are tornadoes that occur over bodies of water. The visible waterspout corresponds to the pendant cloud of the land tornado.

Tornadoes almost invariably occur in the southeast quadrant of an easterly moving cyclonic area and generally move from southwest to northeast at an average rate of from 20 to 50 miles an hour. The path of a tornado is from a few feet to perhaps 2,000 feet in width, while the average length is about 25 miles. Neither the air pressure nor the wind velocity have ever been measured near the center of a tornado but from the force necessary to move certain objects it has been calculated that the wind must blow at the rate of well over 100 miles an hour and may reach several hundred miles.

The tornado tube in its projection downward from the cloud mass is a simple vortex and obeys the laws of fluids in gyratory motion. A partial vacuum is produced at the center of the whirl, the low temperature which results generates the sheath of vapor that makes the tube visible and the wind about the vortex prostrates every obstacle.

Tornadoes may occur in the Gulf States in the winter or early spring, and in the northern States late in the summer. The region of greatest frequency is the central Plains States and the Mississippi Valley, where they occur most frequently in April and May. The southern margin of a tornado is more dangerous than the northern, and as the width of the path of greatest destruction may not be more than a few yards or rods, a person can frequently find safety by run-

ning toward the northwest, if the tornado seems to be approaching directly.

Desert whirlwinds may be tornadic in character or may be due to the convectional overturning of small masses of overheated surface air. The Santa Ana of southern California is caused by cyclonic winds and may last for several days. Other more brief sand storms are due to straight line thundersqualls.

HURRICANES.

Most of the cyclonic storms which gain such a velocity of gyration as to constitute hurricanes originate within the tropics. Those originating north of the equator move northwestward, many reaching latitudes of 20° or more and then recurving toward the northeast. Those of the southern hemisphere first move southwestward, and later, in many cases recurve towards the southeast. Hurricanes are the most destructive of all storms. They have all the characteristics of tornadoes but instead of being a few rods in width, their path of destruction may cover several hundred miles, and instead of their duration being less than one minute as in the case with tornadoes, the terrific winds and rain acompanying them may last from 12 to 24 hours. Hurricanes seldom occur in the northern hemisphere except in the late summer or early autumn. Although there are an average of about 10 annually that touch some portion of the Atlantic or Gulf Coast, an average of less than one a year is severely destructive.

The most intense hurricane of which we have record in the history of the Mexican Gulf Coast, and probably in the United States, moved into the lower Mississippi Valley on September 29, 1915. The pressure fell to 28.11 inches at New Orleans at 5.50 P. M., on the 29th. The wind reached a five minute velocity of 86 miles an hour from the southeast at 5.10 P. M., of the 29th. The extreme velocity was 130 miles an hour. At Burrwood, La., 100 miles south of New Orleans, the velocity was the highest ever recorded on the Gulf Coast. In fact, this was the most intense hurricane known to be recorded in this part of the country. At Burrwood, the extreme wind for one minute was 140 miles an hour, at 3.45 P. M., the maximum five minute velocity was 124 miles an hour, and from 3.31 to 3.50 P. M., the average velocity was 116 miles an hour. From 3.00 to 4.00 P. M., the average velocity was 108 miles an hour, from 4.00 to 5.00 P. M., 106 miles an hour, and from 5.00 to 6.00 P. M., 96 miles an hour. The total loss of life in 300 miles of coast line was only 275. Twenty-three of these fatalities were known to be due to an absolute disregard of warnings, at

Rigolets. The property loss was probably more than $13,000,000. At Leeville, of the 100 houses in the village, only one was left standing.

LAND AND SEA BREEZES.

In warm summer days when cyclonic movements are not well defined, a daily movement of surface winds takes place near the coast analogous to the monsoon winds but of far less extent. The air along the coast flows toward the land in the daytime and toward the water at night. The sea-breeze or daytime wind is felt for only a few miles along the coast but when it does prevail, it furnishes a pleasant relief from the heat. The land and sea-breeze is a *diurnal* wind while the monsoon wind explained in Chapter VIII is a *seasonal* wind.

MOUNTAIN AND VALLEY WINDS.

In some of the mountain regions there is a well-defined movement of the air up the valleys in the daytime and an even more marked movement down the valleys at nighttime.

OTHER WINDS.

Other winds of considerable interest in the United States are the "warm waves," the "blizzard," and the chinook. The first is the warm moisture-laden wind that blows from the south into an advancing cyclonic area. It is particularly marked in the winter time in the central and eastern States, when almost summer heat may be experienced. The Italian name "sirocco" is sometimes given this wind. The blizzard is characteristic of the Great Plains and is a high cold wind accompanied by fine snow or ice particles.

The chinook occurs mainly on the eastern side of the Rocky Mountains particularly in Montana and Wyoming. It is a hot, dry wind which usually makes its appearance suddenly and may raise the temperature 40° to 50° in a few minutes. The snow evaporates very rapidly and large areas previously snow-covered are made available for grazing. This wind has the same characteristics as the foehn wind of the Swiss Alps in fact, "chinook" is merely the local American name for a widespread type of wind to which the generic name "foehn" is now applied by meteorologists.

Various cyclonic and local winds which occur in various parts of other countries are of local interest but are not considered to be of sufficient importance to find a place in this book.

CHAPTER X.

FORECASTING THE WEATHER.

GENERAL CONSIDERATIONS.

It is proposed in this chapter to mention very briefly the preliminary steps in the construction of a weather map, and to offer some suggestions as to the methods of acquiring skill in reading the map and drawing conclusions therefrom as to the weather to be expected in the future.

The history of organized weather services has been often told and will not be repeated here.

The basic material for the construction of a synoptic weather chart is, of course, a sufficient number of meteorological reports from suitably placed stations. It will be assumed that the observing stations have been organized and that the taking of meteorological observations has been begun. By the taking of meteorological observations is meant observing and recording the pressure and temperature of the atmosphere by means of the barometer and thermometer, respectively, determining its moisture contents by the hygrometer, noting the direction and velocity of the wind, the state of the sky with respect to the clouds, the state of the weather at the moment, whether clear, cloudy, foggy, raining, snowing or sleeting, and measuring the precipitation since last observation. It will be further assumed that the reader is not particularly concerned with the reduction of the observation and its preparation for transmission by telegraph or cable to a central station.

At the central station the receipt of the detailed information from a large number of stations calls for systematic treatment in order that it may be made quickly available for use. In all of the various steps leading up to the completion of a weather map, there is an insistent demand for haste that at times militates against efficient work. The first step in the construction of a weather map is, of course, the entering of the data on appropriate maps of the field of observation. An appropriate map is generally understood to be a skeleton map showing the political divisions, the shore lines, the larger rivers and sometimes the most prominent features of the surface relief. The maps also

contain a small circle at each observing station for convenience in expressing the state of the weather and the direction of the wind.

Figure 67 shows a completed weather map, reproduced from a manuscript map such as is made at all forecast centers.

This map shows the presence of a severe cyclone off the New Jersey coast. The barometer level at the center has reached the unusually low level of 28.70 inches while the barometer in the great anticyclone or high that stretches from the Canadian border to the Gulf of

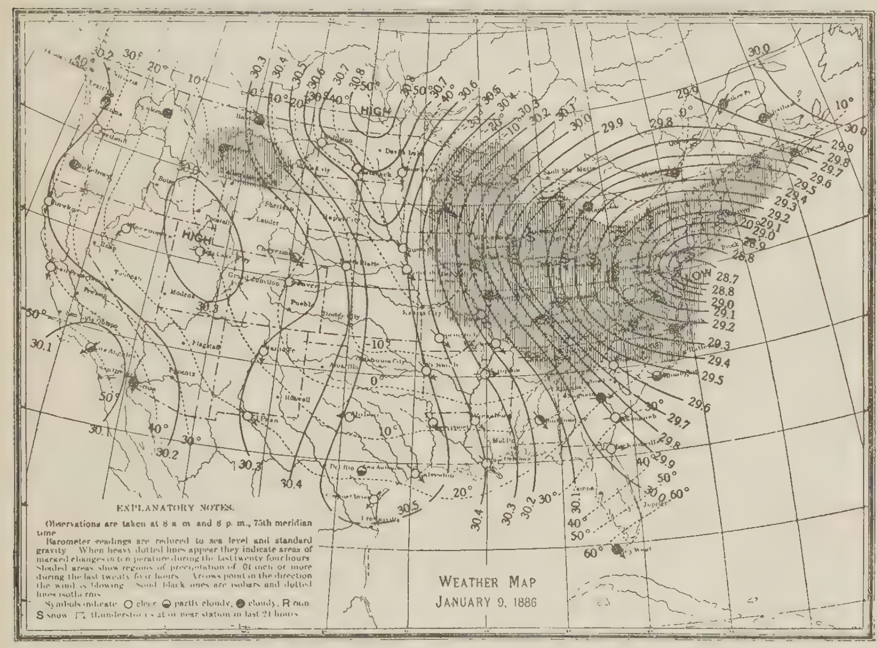

FIGURE 67. Weather Map of January 9, 1886.

Mexico reads 30.50 inches, a difference of nearly two inches of pressure. Great contrasts like this are quite unusual, and while they serve as an illustration of extraordinary conditions the student should not expect to find a similar map in several years' experience with weather maps.

After the data have been entered on the map, the isobars (lines of equal pressure) are sketched in. In drawing isobars one is able to see at a glance the distribution of pressure that prevailed at the moment of observation. It will be noted that pressure is high in some places and low in others; in short, that the isobars wander about much as do

the contours on a topographic map, and, carrying the analogy farther, that there are ridges and hollows in the system of isobars as well as in the contour lines. It is customary to identify the regions where pressure is high by writing the word "High" in the region of maximum pressure and similarly the word "Low" in the region of minimum pressure. The High and the Low of the weather map are synonymous with the anticyclone and the cyclone of the preceding chapter.

Owing to the great importance of the High and Low in the weather forecasting, it is necessary to caution the reader not to consider every barometric configuration marked "High" on the daily weather map as a fully developed anticyclone and the same reservation must be made with respect to the "Low."

As a rule, the words "High" and "Low" are written on weather maps after a preliminary inspection and without a close scrutiny to see whether or not the wind circulation appropriate to anticyclones and cyclones, respectively, is present in each case.

The art of weather forecasting rests almost entirely upon the fundamental proposition that weather travels. If, therefore, the field of observation is sufficiently extensive and the character of existing weather is accurately portrayed on appropriate charts, usually there is not much difficulty in announcing the weather to be expected for any place in the line of travel. Practically all of the rules known and used in the art have been established empirically; some of them have been formulated, but in a considerable proportion of cases the rules which govern the forecaster are exercised subconsciously. The difficulty in formulating any set of rules for the making of weather forecasts is further increased by the fact that an appropriate nomenclature of the art does not exist. Without nomenclature it is difficult to write even for those who have a technical knowledge of the subject.

The writer has elsewhere stated that the two fundamental propositions of weather forecasting are (1) weather travels, and (2) the weather for any time and place depends almost wholly on the pressure distribution that obtains for that time and place, hence the first and most important thought of the forecaster is to form a mental picture of the pressure distribution 24 to 36 hours in advance. This determined with a fair degree of accuracy, it is not difficult to sketch in the appropriate weather.

The problem of forecasting the weather is not without its difficulties and should not be lightly approached, nor should the hope be entertained that it can be mastered in a few weeks or a few months. True, the broader rules can be given in a few hours but the application of the rules presents difficulties of no small order.

The student will soon discover that the weather is in a perpetual condition of change; in other words, it would seem that in nature an effort is made automatically to reach a state of equilibrium, a condition, needless to say, that is approached but never reached.

The taking of observations and the making of weather maps is a continuous process; seven hundred and thirty maps are made in the course of a year when but two regular observations are made daily. Counting special observations, close to a thousand maps are made each year by the United States Weather Bureau. No two of these maps are precisely alike, but just as some human faces differ only in a slight degree so some weather maps have a close resemblance. On close scrutiny, however, points of difference will be found that are so ill-defined as to escape notice on casual inspection. Many investigators have urged the classification of weather maps by types and indexing the types for ready reference, the object being, of course, to make a forecast based upon what followed the original type map. The writer does not discourage the classification of weather maps by type; on the contrary, the work of classification, if carefully done, will establish firmly the fact that no two maps are precisely alike in the lesser details and even in some of the greater details. Similar maps may be found for some individual date and hour, but it has been the experience of the writer that the similarity diminishes directly with the passage of time. In other words, the subsequent maps will diverge more and more from the original, and frequently in a different direction.

If one carries out a plan of classifying maps by types for even a short time the point will be reached when the number of types will become unwieldy and there will be a rather large number of maps that do not easily fall into the adopted classification. However, the study of types is not without its advantage. To illustrate, suppose the object in hand is to forecast heavy snow for a given locality. The procedure would be to list all of the cases of heavy snow that had occurred in the past and then assemble the weather maps for the corresponding dates 24 to 36 hours in advance of the snowfall. It can then be seen whether or not heavy snow for the locality occurred with a certain type of map or promiscuously with several types of map.

FORECASTING THE WEATHER AND THE TEMPERATURE.

The great bulk of the work of the forecaster is in preparing twice daily, morning and evening, forecasts of weather, wind and temperature. This branch of the work is the most unsatisfactory because

of the fact that the indications of the map are often obscure and ill-defined. Often the maps are neutral, that is, they give no positive indications.

The weather usually travels from west to east. But we are not to consider the movement of cyclones and anticyclones, or the Lows and Highs of the weather map, as of the same order of certainty as that of a railroad train even under the unfavorable conditions which sometimes obtain on railroads. In the beginning, therefore, the student will do well to locate definitely the positions of both Low and High 24 and 36 hours in advance as best he can. The average paths of Lows which enter the United States in January and July, as well as the average rate of progression, is shown in Figures 68 and 69 below.

Remarks on Figures 68 and 69.—The paths of cyclonic depressions passing across the United States for a period of 22 years is shown in Figures 68 and 69. The depressions have been classified in accordance with their place of origin; thus, all depressions first observed in the Province of Alberta or adjoining provinces have been classified under the type "Alberta"; those moving inland from the North Pacific have been classified under the type "North Pacific"; in like manner practically all the depressions that may move over the country have been grouped in classes. The average daily movement in each class is also shown. The fact that certain definite tracks have been deduced for certain groups of storms should not be taken too literally. As a matter of fact, there is no portion of the United States east of the Rocky Mountains that at one time or another is not passed over by cyclonic depressions. The width of a depression may vary from a minimum of, say, 300 miles, to a maximum of 1500 miles, but manifestly the charting of these storms must be confined to the central portion and it is the central portion that is represented by the paths shown on Figures 68 and 69. The percentage of cyclonic depressions from the different districts is as follows:

Alberta	38%
North Pacific	15
South Pacific	7
Northern Rocky Mountain	5
Colorado	12
Texas	9
East Gulf	3
South Atlantic*	3
Central	7

* The expression "South Atlantic" refers to the waters of the North Atlantic bordering the southeastern United States and north of the West Indies. Simi-

The chief characteristics of a Low are cloudy skies in and for some distance in front of its center, from which rain falls intermittently as in the showers of summer, or steadily as in the rain of the cold season. The distance in front of a Low at which precipitation may occur is anywhere from a few miles to a thousand miles. The maps do not afford any indication as to how far in front of the Low precipitation may occur. In the upper Missouri Valley and thence westward to the Pacific precipitation occurs mostly in the rear of Lows or, what is almost the same thing, on the front of the High which follows immediately in the rear of the Low. There is no sharp line of demarcation between the western side of a Low and the eastern side of a High when they are moving in tandem formation across the country. The isobar of thirty inches is sometimes taken as separating the Low from the High but there is no fixed rule. Reference to Figure 67 will show that the character of the weather on either side of the isobar of 30 inches is not markedly different and that it is not until a great distance from the cyclone center is reached that the characteristic cloudiness of the cyclone disappears.

It will be found that the speed of the Low is a factor in estimating the probability of precipitation. Much more rain falls in a Low that moves slowly than in one that moves rapidly. The level of the barometer at the center of a Low is also a factor in determining the probability of precipitation although there is no rule of general application throughout the year.

Snow falls apparently with less effort than rain and with a higher barometer in the Low. Heavy snows have been observed with Lows in which central pressure was as high as 30.20 inches. Rain does not readily fall with central pressures above 30.00 inches.

The average rate of movement of Lows may be seen from Figures 68 and 69. It is suggested that the student make it a regular practice to forecast each day the occurrence or non-occurrence of rain and that he save the daily weather maps immediately prior to the dates on which precipitation occurred, classify them and analyze his failures in the light of subsequent events. In no other way can skill be acquired in reading weather maps so far as the local situation is concerned. Forecasting temperature changes is less difficult than forecasting the occurrence of precipitation. In forecasting temperature changes one has to

larly, "North Pacific" and "South Pacific," as here used, refer to the northern and southern parts, respectively, of the North Pacific Ocean adjacent to the United States.

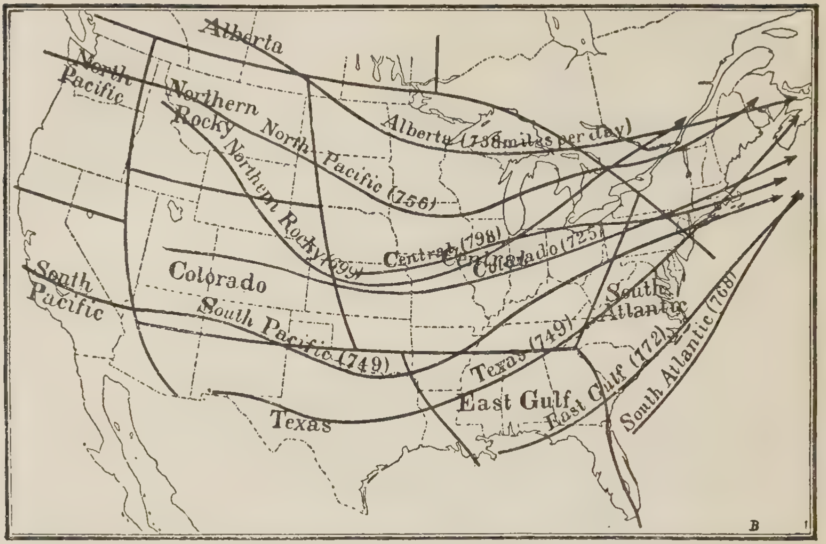

FIGURE 68. Average Paths and Daily Movement of Lows in the United States, January.

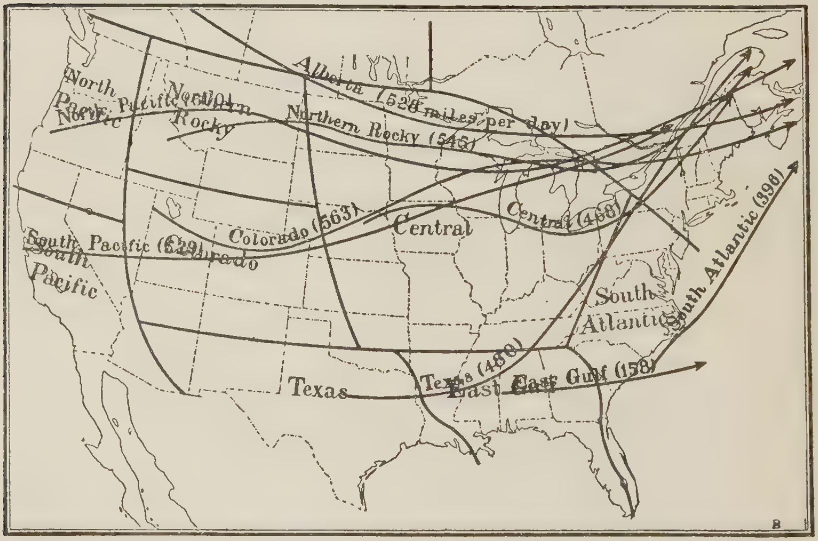

FIGURE 69. Average Paths and Daily Movement of Lows in the United States. July.

keep in mind the fact that the atmosphere is heated mostly by radiation and reflection from the earth and that there is superposed upon the diurnal system of heating the atmosphere in lower levels what may be called the cyclonic system whereby warm air from low latitudes is transported to higher latitudes and vice versa. This transportation is brought about by the movement eastward of cyclones. The temperature of a place is alternately elevated and depressed with the passage of a cyclone and the forecasting of warmer and colder is an easy matter provided the center of the low pressure system is accurately foreseen. And again we are reminded that the key to successful forecasting lies in the ability to locate the center of the Low 24 and 36 hours in advance. Several chapters could be written upon the perturbations of Lows and Highs but it is doubtful if the recital of them would be helpful.

As bearing upon the movement of Lows and Highs the following observations may aid the beginner. It is important to note carefully the movement of both the Low and the two Highs which attend it—one on the front and the other in the rear, since the movement of the High in front of the Low seems to have an important bearing upon the movement of the latter. This is equivalent to saying that the pressure distribution should be carefully considered. If the High in front is moving rapidly the Low in its rear will almost invariably have a rapid movement also. If pressure is low in the probable path of advance the movement is apt to be rapid; in fact, there are occasions when one Low appears to merge completely with another Low immediately in its front.

The barometer level in the center of both Highs and Lows should be carefully watched. Rising pressure in the center of a Low is an indication that the Low is diminishing in intensity, and conversely, rising pressure in the center of a High is an indication of increasing intensity in that formation, whereas if the pressure at the center of a Low is decreasing it is a sure sign that the intensity of the Low is increasing.

The trend of the isobars is often significant, since it gives information as to the direction of the winds; thus isobars trending in a north-south direction as in Figure 67 (see the region west of the Mississippi), indicates that there will be a transfer of cold air from high to low latitudes on the eastern side of the High and consequently much colder weather will prevail, while on the west side, the flow will be from south to north with higher temperature on that side. A High in which the

longer axes extends in an east-west direction is generally attended by clear skies with west winds on its north side and frequently by considerable cloudiness with east-northeast winds on its south side.

Rain Forecasts.—According to theory, partly if not fully supported by observation, there is a small component of motion directed upward in a cyclone and downward in an anticyclone. That the air ascends in a cyclonic system cannot be doubted but that the air descends in an anticyclone is not always obvious. The forecaster in the beginning of his work must naturally assume that precipitation will occur in connection with each cyclone that appears within the field of observations. He may feel some doubt in those cases where the sky remains clear in spite of the cyclonic circulation. In the interior of a continent, far removed from the source of moisture, clear skies at the center of a cyclonic system may be expected, but when the sky remains clear as the cyclone approaches a source of moisture, there is reason to hesitate before predicting precipitation. One of the paradoxes of meteorology is the existence of cyclones without condensation of moisture even sufficient to cloud the sky. Cyclones without precipitation are the exception, however, and the beginner may well defer consideration of them for the present.

The relative position of the Low with respect to the Highs which accompany it is worthy of a moment's consideration. The ideal situation of a Low as regards its future maintenance and development is that one which places it between two Highs, one in front and the other in the rear. This situation frequently occurs in the cold season when cyclonic storms move northeastward from, say, the lower Mississippi Valley. In a configuration of this sort, the eastern High serves as a reservoir from which air is drawn into the cyclone on its eastern side, ascends in the cyclone; its moisture is condensed and the cooled air then flows off laterally, probably merging with the general eastward drift of the air at the level at which it finds itself after cooling and later descends in an anticyclone many miles from the place of its ascent. Forecasting of rain in this case involves no difficulties other than in determining the cyclone's path and how far in advance of the cyclonic center precipitation will occur in the ensuing 24 hours. Obviously there are two conclusions to be reached, first as to the position of the cyclone itself and second as to the distance in advance of the center that precipitation will occur. Much of the uncertainty of rain forecasts rests in the inability of the forecaster to determine the eastern limit of the rain front and how long precipitation will continue after a beginning has been made.

Forecasting Temperature Changes.—The atmosphere is heated mainly by contact with the earth's surface and the conditions which determine its temperature are therefore mostly the same as those which determine the temperature of the latter. The character of the surface whether land or water and if the former, whether bare or covered with vegetation, is therefore an important consideration. The forecaster is concerned chiefly with the transport of warmer or colder air from one point to another by the circulation induced by traveling cyclones and anticyclones. If a cyclone is approaching in the cold season and there will be an inflow of southerly winds, warmer weather will obtain in front of the cyclone; the amount of the warming and the extent of territory over which it is effective are details that can best be learned by experience. Temperature forecasts are free from the perplexities that are associated with rain forecasts.

Forecasting Strong Winds, Cold Waves, etc.—Roughly speaking, the strength of the wind is proportional to the barometric gradient. The word gradient is used in the same sense that an engineer would employ it to indicate the slope from a system of contour lines. The barometric gradient is susceptible of quantitative determination; thus, let the point for which the gradient is desired be located within a system of isobars. Draw a line through the point at right angles to the isobars on either side and measure the distance along the line from high pressure to low. If the distance be 100 miles and the difference in pressure 0.1 inch, then the gradient per mile will be $\frac{0.1}{100} = 0.001$ inch of mercury per mile. In practice the arithmetical calculations are seldom made but it is customary to use the distance between isobars as the inverse of the gradient; thus when the isobars are close together as about the Low center in Figure 67 the gradient is strong; when they are wide apart as in the region of the High west of the Mississippi, the gradient is weak. In the region around the center of the Low, Figure 67, the gradient is unusually strong and fresh gales should prevail there; in the region of the High, the winds should be gentle.

In forecasting the winds about a Low, therefore, the forecaster must consider the probability of the gradient remaining constant, increasing or decreasing. If the probable path of the Low will bring it into a region of uniform temperature distribution, the probabilities are that the Low will decrease in intensity, if into a region of great contrast in temperature the Low will increase in intensity.

The geographic position of the Low is also a factor in determining

whether or not it will increase in intensity. Ordinary Lows increase in intensity when they move from lower to higher latitudes, particularly along the Atlantic coast.

Every cyclone seems to contain the possibility of a marked increase in intensity as it nears a coast line. On the Pacific coast storms rarely retain much of their original strength when passing inland, but along the Gulf of Mexico and the Atlantic coast, the original severity of the storm is often retained for a short time after it passes inland. Cyclones moving northeastward along the Atlantic coast generally increase in intensity.

The best guide as to whether or not a cyclonic system is or is not increasing in intensity, is the rate of pressure fall immediately in advance of its probable path.

Weather Bureau observers are required to include in their telegraphic reports of observations a word to show the amount of the rise or fall of the barometer in the 2 hours immediately preceding the observation. A fall of as much as 0.07 inch in an hour or 0.14 inch in 2 hours indicates a decided increase in the intensity of the cyclonic system.

The cold waves of winter are also conditioned upon the occurrence of marked disturbances in the pressure distribution, such as illustrated in Figure 67. That particular map is one of a series that illustrate the conditions which preceded the very severe cold wave of January, 1886, when zero temperature was reached as far south as northern Georgia.

The line of zero temperature on the map of the 9th has reached northern Mississippi (see the isotherms—dotted lines).

Seasonal Influence.—The change of weather with the season is of course well known. Winter is the season of sharp temperature contrasts and great storm activity. The movement of disturbances is at a maximum and the problem of determining the pressure distribution on a succession of maps is most difficult. In summer the horizontal circulation is at a minimum and convective circulation is at a maximum. The continental interior is then warmer than the adjacent oceans and the forecasting of pressure distribution in summer is very different from that of winter. In summer the weather controls are largely local rather than general. Thunderstorms and convective rains are the rule, especially in the southern part of the United States. While the barometric configurations aid the forecaster to a certain extent, considerations of local heating, the prevailing winds, and the moisture content of the atmosphere are also employed.

THE WEATHER IN AVIATION.

Improvement in airplane construction, the development of high-grade engines, etc., etc., has made the aviator more or less independent of weather conditions, save those of fog and cloud. High winds if not gusty do not appear to be a hindrance to successful and prolonged flight but the presence of fog or a cloud layer which obscures the vision is a serious menace. The weather map when properly interpreted contains useful information for the aviator. Much useful information has also been secured in recent years in the exploration of the free air by means of kites and balloons. In this category must be included our knowledge of the shifting of the wind with altitude. The law expressing the deflection of the wind to the right with increase of altitude was announced more than 50 years ago, from observations on the movement of clouds. More recent determinations by the use of kites and balloons show that the turning of the wind is quite definitely associated with the pressure distribution for the time and place, hence an intelligent application of the data of the weather map will put the aviator in possession of important information as to the probable changes in wind direction aloft. The condition of the weather as to the prevalence of cloud can be inferred, remembering that the center of a High is a region free from cloud, while a Low is a region of cloud and rain, particularly on its eastern front.

The High is more enduring than the Low and the movement is naturally slower, hence when a High sets in, settled weather for a day or so, sometimes longer, may be anticipated.

It is quite important that the aviator recognize signs of the approach of falling weather. If for example the weather map shows a cyclonic depression approaching from the west the formation of a cloud veil and the freshening of the wind should enable him to interpret the coming weather with some certainty. In the absence of these preliminary signs it may be assumed that the depression has not yet begun to influence the weather.

Kite flights in the United States have shown that easterly winds are infrequent and generally of low speed, also that they do not extend upward much more than 3 kilometers and rarely even to that elevation. Easterly winds are more frequent on the eastern coast than inland. In the interior of the continent, however, easterly winds may be expected in connection with an anticyclone moving southeast or east. The easterly winds will be found, of course, on the southeastern and southern margin of the anticyclone and they may be expected up to a

height of about 3 kilometers. Over the central region of the anticyclone, however, the winds will be, as a rule, from the west at an altitude of 4 kilometers and upward. The one outstanding fact with relation to the wind aloft in the United States is that it is uniformly from the west at all seasons, having a small northerly component in the cold season. The wind in connection with cyclonic depressions is quite uniformly from the southwest on the eastern side of the depression and from the northwest on the western side. The elevation to which southwest winds extend is variable and it is not improbable that the intensity of the cyclone is in some way dependent upon the depth of the southwest wind on its front. Observations show that the altitude of the southwest wind on the eastern front of cyclonic depressions ranges from 2 to 3.5 kilometers; doubtless it extends higher but probably the kites have not as yet penetrated to the top of the southwest stratum. The motion of cirrus clouds is frequently from the southwest over the interior of the country. This fact would seem to indicate that the southwest wind has a much greater altitude than can be inferred from kite flights.

CHAPTER XI.

CLIMATE.

DEFINITION OF CLIMATE.

THE *climate* of a place or region is the complex of weather conditions to which it is subject, as inferred from past experience. The description of a climate includes an account not only of average and typical conditions, but also of the range of the meteorological elements and the frequency with which extreme and abnormal conditions occur. The branch of meteorology dealing with climate is called *climatology*. As distinguished from discussions of climate in general, descriptions and statistics of the climates of the world constitute a phase of climatology which is sometimes known as *climatography*. Climatographic works on particular regions are, however, also called "climatologies"; e. g., Henry's Climatology of the United States.

"Climate" and "weather" are both terms which, as commonly used, refer especially to those meteorological phenomena having important direct effects upon living organisms and upon the affairs of humanity. Temperature is the leading element of climate. On the other hand, barometric pressure, so prominent in meteorology, is almost ignored in climatology, because it is of little immediate interest from either the biological or the economic standpoint.

CLIMATIC STATISTICS.

One of the leading tasks of the climatologist or climatographer is to collect the data of observations made in the course of years at meteorological stations and digest them in the form of climatic statistics. Such statistics are utilized in many of the arts and sciences, including geography, agriculture, biology, medicine, etc. There is a striking disparity in the amounts of climatic information available for various parts of the world. Thus, the climatography of the ocean, which lacks fixed meteorological stations, is less well known, on an average, than that of the land. Climatic statistics are fairly abundant for western Europe, the United States, Canada, Australia, British India, Japan and the temperate regions of South America. They are very scarce for the

uninhabited polar areas, the uncivilized parts of Asia and Africa, the Ottoman Empire, and many Latin-American republics.

One fact, however, should be made clear, and it has not previously been emphasized in works on meteorology: There is no part of the world for which the existing climatic statistics are adequate to meet the diverse demands constantly made for such data. This is partly due to the limited scope of meteorological records; but even the records that exist have been digested only in a very fragmentary way for the purposes of climatology. The field for valuable work in this line is unbounded, and is commended to the attention of the young student.

There is a diversity of opinion as to what kinds of data should be included among the statistics of climate, and those actually presented vary widely. The accompanying table of so-called "normals" and extremes for Washington, D. C., illustrates the practice of American climatologists in the form and selection of data. (For practical purposes, a normal is an average based on a record extending over a long series of years. The data tabulated herewith are from the following periods of observation: Mean maxima and minima, 32 years; humidity, 15 years; sunshine, 14 years; all other data, 33 years.)

In the first column we have the mean temperatures of the months, the seasons, and the year. These are obtained by averaging the corresponding data for the individual years of the record. The monthly means for the individual years are obtained by averaging the daily means, which, in turn, are the means of the daily maxima and minima.[1] The mean annual temperature is the mean of the twelve monthly temperatures. Similarly, the seasonal means are obtained by averaging the three monthly values for the appropriate seasons.

The second column shows the mean daily maximum temperature. This should be distinguished from the mean monthly maximum temperature, which is not shown in the accompanying table but is a datum of importance, and is often given in foreign climatic tables. The first figure in the column, 44, means that on an *average* day in December the thermometer at Washington rises only to 44° F. The mean monthly maximum would show the highest individual reading of the thermometer recorded during an average December.

[1] The true mean daily temperature is the mean of 24 hourly observations. This is obtained only approximately by taking the mean of the maximum and minimum. In European practice it is often obtained by applying certain formulæ to the readings made at two or more specified hours. See Hann's Handbook of Climatology, tr. by Ward, p. 7-8.

Climatological Data for Washington, D. C.

(From Henry's "Climatology of the United States.")

Latitude, 38° 54′ N. Longitude, 77° 3′ W. Elevation, 73 feet.

Month.	Temperature. Mean. °F	Mean of the maxima. °F	Absolute maximum. °F	Mean of the minima. °F	Absolute minimum. °F	Highest monthly mean. °F	Lowest monthly mean. °F	Precipitation. Mean. In.	Number of days with 0.01 or more.	Total amount for the driest year. In.	Total amount for the wettest year. In.	Snow. Average depth. In.	Snow. Greatest depth in 24 hours. In.	Mean humidity. Relative, 8 a. m. P.ct.	Absolute, 8 a. m. Grs.	Relative, 8 p. m. P.ct.	Absolute, 8 p. m. Grs.	Total sunshine. Average hours.	Percentage of possible.	Direction of prevailing wind.
December	36	44	73	29	—13	46	26	3.1	10	4.2	0.2	2.8	10.0	76	1.66	67	1.77	158	54	NW.
January	33	41	76	26	—14	44	25	3.4	12	2.1	4.0	5.9	5.6	77	1.49	69	1.63	145	48	NW.
February	35	44	78	27	—15	43	26	3.6	10	4.6	2.5	8.1	12.0	75	1.52	66	1.62	151	50	NW.
Winter mean	35	43	..	27	..	..	..	10.1	32	10.9	6.7	16.8	..	76	1.56	67	1.67	151	51	NW.
March	42	51	83	33	4	50	34	4.1	12	1.0	4.2	4.6	10.0	75	1.98	65	2.14	179	48	NW.
April	53	63	93	43	22	58	48	3.2	11	3.3	9.1	0.4	4.0	70	2.76	59	2.96	211	53	NW.
May	64	74	96	54	34	70	59	3.8	12	4.0	10.7	0.0	0.0	75	4.31	68	4.61	244	55	S.
Spring mean	53	63	..	43	..	..	..	11.1	35	8.3	24.0	5.0	..	73	3.02	64	3.24	211	52	NW.
June	73	83	102	63	43	78	67	4.0	10	1.2	5.0	0.0	0.0	76	5.87	71	6.24	279	63	S.
July	77	87	103	68	52	81	72	4.5	11	2.1	8.1	0.0	0.0	77	6.76	72	6.95	291	64	S.
August	75	84	101	66	49	80	72	4.0	11	2.0	3.1	0.0	0.0	80	6.81	74	6.92	276	65	S.
Summer mean	75	85	..	66	..	..	..	12.5	32	5.3	16.2	0.0	..	78	6.48	72	6.70	282	64	S.
September	68	78	104	59	38	77	62	3.5	8	1.5	3.9	0.0	0.0	81	5.31	76	5.68	252	68	S.
October	57	66	92	47	26	63	51	3.1	9	3.1	4.5	T.	T.	80	3.50	72	3.61	208	60	NW.
November	45	54	80	37	12	51	40	2.8	10	1.5	6.0	0.7	2.5	79	2.33	69	2.44	154	51	NW.
Autumn mean	57	66	..	48	..	..	..	9.4	27	6.1	14.4	0.7	..	80	3.71	72	3.91	205	60	NW.
Annual	55	64	104	46	—15	..	..	43.1	126	30.6	61.3	22.5	12.0	77	3.69	69	3.88	212	57	NW.

The third column gives the absolute maximum temperatures—i. e., the highest readings ever recorded—for each month during the whole period of observation. A more extensive climatic table would specify the years in which these extreme values were observed, and perhaps also the dates of occurrence. The entry 104 at the foot of the column is the "record" high temperature for Washington during the 33 years covered by the observations.[2]

The fourth and fifth columns give mean daily minimum and absolute minimum temperatures respectively. These are analogous to the maximum temperatures just explained.

The sixth and seventh columns give the highest and lowest monthly mean temperatures, selected from the values for the individual years on which the data in the first column are based.

Under the heading "Precipitation" the first column shows the average total precipitation in inches, for each month, each season, and the year. Snow and other solid forms of precipitation are included here, as well as rain, and are expressed in their "water equivalent."

The next column gives the average number of "rain-days" or days with a measurable amount of rain. The minimum amount of precipitation that must be reached or exceeded to constitute a rain-day is defined in the United States and Great Britain as 0.01 inch, but various other limits are used elsewhere. In countries using the metric system, 0.1 millimeter (0.004 inch) is frequently adopted. Hence statistics of the number of rain-days for various parts of the world are not comparable.

The next two columns require no explanation.

Under the heading "Snow" we have, first, the average total snowfall, as actually measured (not reduced to "water equivalent"), and next, for each month, the greatest actual snowfall in 24 hours ever recorded during the period of observation.

The two columns headed "Total sunshine" show the average duration of bright sunshine, from corrected readings of a sunshine-recorder, for each month, etc., and the average ratio, in percentage, of the actual sunshine to the maximum possible amounts for the respective periods.

The last column shows the prevailing (i. e., most frequent) direction of the wind. This datum, in the case of records of the United States Weather Bureau, is based upon observations, to eight points of the compass, at the regular observation hours (now 8 A. M. and 8 P. M., 75th meridian time). At stations where the wind shifts more

[2] This record was broken on August 6, 1918, with a temperature of 106° F.

or less regularly with the time of day—for example, where alternating land and sea breezes prevail—data based on more frequent observations or on continuous automatic records would be preferable.

Of the numerical data not included in the accompanying table, but often found in climatographic literature, the following are among the more important:

Mean interdiurnal variability of temperature. (Mean difference between mean temperatures of successive days.)
Average and extreme dates of first and last frost.
Ground temperatures at various depths.
Data of relative and absolute humidity.
Data of excessive precipitation (especially the maximum in 24 hours).
Frequency of long dry and rainy periods. (Various specifications for such data have been adopted in different countries. A recent drought-chart published by the U. S. Weather Bureau defines a drought as a period of 30 or more consecutive days during which there is no rainfall amounting to 0.25 inch in 24 hours. In England an "absolute drought" is defined as a period of more than 14 consecutive days without 0.01 inch of rainfall on any day, and a "partial drought" as a period of more than 28 consecutive days, the mean rainfall of which does not exceed 0.01 inch per day. English meteorologists also define a "rain-spell" as a period of more than 14 consecutive days every one of which is a rain-day.)
Number of days with snow, hail, thunderstorms, fog, gales, etc.
Data of cloudiness. (Average number of clear, partly cloudy and cloudy days, or average degree of cloudiness on a scale in which 0 = cloudless and 10 = completely overcast.)
Data of wind velocity. (In miles per hour or meters per second, if from anemometer readings; otherwise on the Beaufort or other wind-scale.)
Phenological data.

Numerical data of climate are supplemented by text descriptions and also by data presented in the form of charts, graphs, wind-roses, etc. On climatic charts are employed various "isograms" similar to those used in presenting instantaneous conditions on synoptic weather maps. The most important climatological isograms are *isotherms,* used in charting temperature, and *isohyets* (generally combined with tinting or shading), used in charting precipitation.[3]

FACTORS THAT CONTROL CLIMATE.

If the earth had a smooth surface, all land or all water, and no atmosphere, the distribution of solar heat would be purely a question of latitude, and we should have a regular gradation of climates, from

[3] For a complete list of the isograms used in meteorology and climatology, see Monthly Weather Review, Wash., April, 1915, p. 195-198.

the hottest at the equator to the coldest at the poles. Although this ideal arrangement, sometimes described as *solar climate,* is profoundly modified by the topography of the earth and the effects of atmospheric and oceanic circulation, it remains true that latitude is the most important single factor controlling climate, with respect to temperature. The effect of altitude upon temperature is analogous to that of latitude, and is seen in the prevalence of cool climates at high levels near the equator as contrasted with the mild climates of many places near sea-level in the temperate zones.

The distribution of land and water, combined with the effects of prevailing winds, gives us an important classification of climates with respect to *range* of temperature. A region in which the winds blow from an adjacent ocean has, for its latitude, mild winters and cool summers, and a comparatively small variation of temperature between day and night; a region not exposed to such winds exhibits a greater range of temperature, both annual and diurnal. In one case the climate is said to be *marine;* in the other *continental.* For example, the western seaboard of Europe, where westerly winds predominate, has a marine climate, while the interior of Asia, which is not only remote from the ocean but also cut off from ocean breezes by mountain barriers, presents an extreme example of the continental climate.

Warm and cold ocean currents necessarily influence to some extent the temperature of adjacent regions to the leeward of them. While there has been a popular tendency to exaggerate this influence, scientific opinion has sometimes gone to an extreme in the opposite direction.

The wetness or dryness of a climate is determined especially by the prevailing movement of moisture-bearing winds and the relief of the land, while a second and important control is the location of a region with respect to storm-tracks. The rainiest regions of the world are found on the windward slopes of mountain ranges not far from the ocean, where the moist winds, forced by the mountains to ascend rapidly, cool dynamically and shed their moisture. To the leeward of such mountains arid conditions generally prevail. Thus the southern slopes of the eastern Himalaya receive an enormous rainfall from the southwest monsoon, blowing from the Indian Ocean, and an abundant rainfall also prevails on the south slopes of the high mountains of eastern Tibet, while northern Tibet is a desert. Rainfall due to the relief of the land is called *orographic rainfall.* Regions not exposed to such rains may nevertheless have a copious rainfall due to the frequent passage of cyclonic storms (*cyclonic rainfall*) or to the frequent occurrence of thunderstorms. The rainfall of the eastern

United States is mainly cyclonic (including the rain falling in thunderstorms of cyclonic origin), while over a considerable part of the equatorial regions the rainfall is mainly due to thunderstorms not connected with cyclones.

CLASSIFICATIONS OF CLIMATE.

Some of the broad classifications of climate have been indicated in the foregoing section. There are, however, many others, dependent not only on the particular element of climate emphasized therein, but also upon the standpoint of the classifier. Thus there are classifications from the point of view of the agriculturist, the medical man, etc. The accompanying table, prepared by Dr. W. F. R. Phillips, is a convenient synopsis of the classifications in common use. To these should be added a classification of annual rainfall, mainly with reference to biological effects, originally proposed by F. Waldo, viz.:

Excessive rainfall, over 75 inches.
Copious rainfall, 50-75 inches.
Moderate rainfall, 25-50 inches.
Light rainfall, 10-25 inches.
Desert rainfall, under 10 inches.

Principal Classifications of Climate.

(Adapted from A. H. Buck's Reference Handbook of the Medical Sciences.)

Classification Basis.	Subdivisions Under Classification.	General Characteristics of Each Subdivision.
Solar or astronomical.	Tropical	Usually mild, equable, moist, warm, average temperature 80° F. Rainfall frequent and heavy over water and over windward land exposures. Nights usually clear, afternoons cloudy. No general storms. Seasons, rainy and dry; but this division is only a relative one.
	Temperate	Unsettled weather, great and variable changes in temperature, rainfall, and moisture from season to season and day to day. Region of cyclonic storms, cold and hot waves, floods and droughts.
	Polar	Cold; temperature on average considerably below freezing. Scanty rainfall. Very short but hot summer. Winter long and severe. Storms infrequent.

Classification Basis.	Subdivisions Under Classification.	General Characteristics of Each Subdivision.	
Geographical.	Continental	High temperature in daytime, low at night. Difference between day and night temperatures increases toward center of continent. Great variations in temperature, sometimes hot, sometimes cold. Moisture variable from almost saturation to aridity. Rainfall subject to great variations and extremes. General tendency to extremes in all climatic elements.	
	Marine	Temperature equable; range between day and night hardly exceeds 2° to 4° F. in mid-ocean. Moisture high but constant. Rainfall frequent.	
	Insular and littoral	Intermediate between above, partaking more or less of the characteristics of one or the other.	
Topographical (land).	Plain	Extremes of temperature great, rainfall uncertain, humidity low.	
	Hill	Extremes of temperature less than plain, rainfall greater, humidity higher.	
	Mountain	Generally same as hill, except effects of altitude become more evident; rainfall increases up to about 5,000 feet, then decreases.	
	Valley	Extremes of temperature greater than hill. Humidity greater; rainfall greater than plain.	
	Desert	Nearly rainless; great extremes of temperature between night and day and season and season.	
	Artificial	Such as climates of large towns. Mean temperatures generally higher in large towns than in surrounding country, but ranges of temperature somewhat less; haze, cloud and fog more frequent.	
Aërophysical.	Temperature	Hot Intermediate Cold	According to the degree of heat adopted as the standard of comparison.
	Humidity	Damp or moist Intermediate Dry or arid	According to standard of humidity adopted.

Classification Basis.	Subdivisions Under Classification.	General Characteristics of Each Subdivision.
Physiological.	Invigorating ...... Relaxing .... Rigorous, etc.	According to the general effects of the particular climate.
	Mild .... Pleasant .... Humid .... Disagreeable, etc.	According to the general sensation produced, etc.

CLIMATIC ZONES AND PROVINCES.

The common geographic division of the earth's surface into the torrid, temperate and polar zones, based on astronomical considerations and bounded by the tropics and polar circles, is somewhat misleading for the purposes of climatology. In so far as this subdivision has reference to the distribution of temperature, it is better to limit the zones by isotherms, annual or otherwise, as has been done by Alexander Supan in the chart shown herewith.

Since political boundaries are not laid down on the basis of climate, the climatologist finds himself embarrassed by the common question, "What is the climate of such and such a country?" A prospective emigrant will, for example, write to the Weather Bureau to ask what kind of climate he may expect to find in South America or Australia. A diversity of climates is, of course, found within the confines of every large country and most small ones. It is, however, possible to delimit areas in which the climate is fairly homogeneous and is therefore well represented by the statistics of a single meteorological station situated therein. This conception of climatic units or *provinces* is an important one.

A division of the whole surface of the earth into thirty-five climatic provinces, based principally upon temperature and rainfall, but partly, also, on winds and orographical conditions, is shown in the accompanying chart, by Supan. More refined classifications have been proposed for particular countries. These are exemplified, in a measure, in the so-called "forecast districts," adopted in most countries having meteorological services.

Much important work remains to be done in the delimitation of climatic provinces. Indeed the whole fruitful subject of comparative climatology has been seriously neglected.

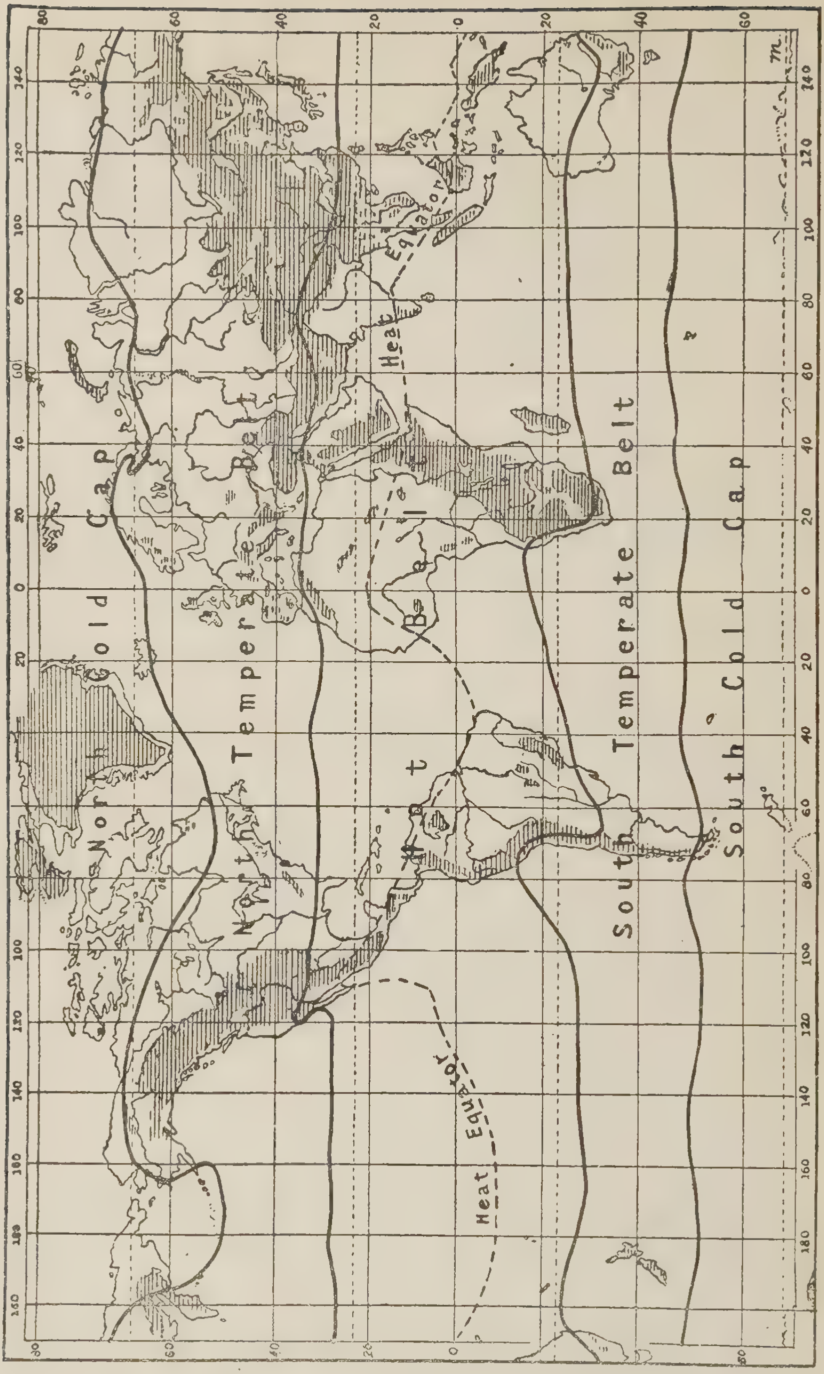

FIGURE 70. Supan's Temperature Zones. (From Bartholomew's Physical Atlas.) The plateau area above 3,000 feet is shown by shading. The Heat Equator, or line of maximum mean annual temperature, is shown by the dotted line. The Hot Belt is bounded to north and south by the isotherm representing the mean annual temperature of 20°C. (68°F.). The Temperate Belts lie between these lines and the isotherm of 10°C. (50°F.) for the warmest month. The cold Caps cover the regions around the poles to the isotherm of 10°C. (68°F.) for the warmest month.

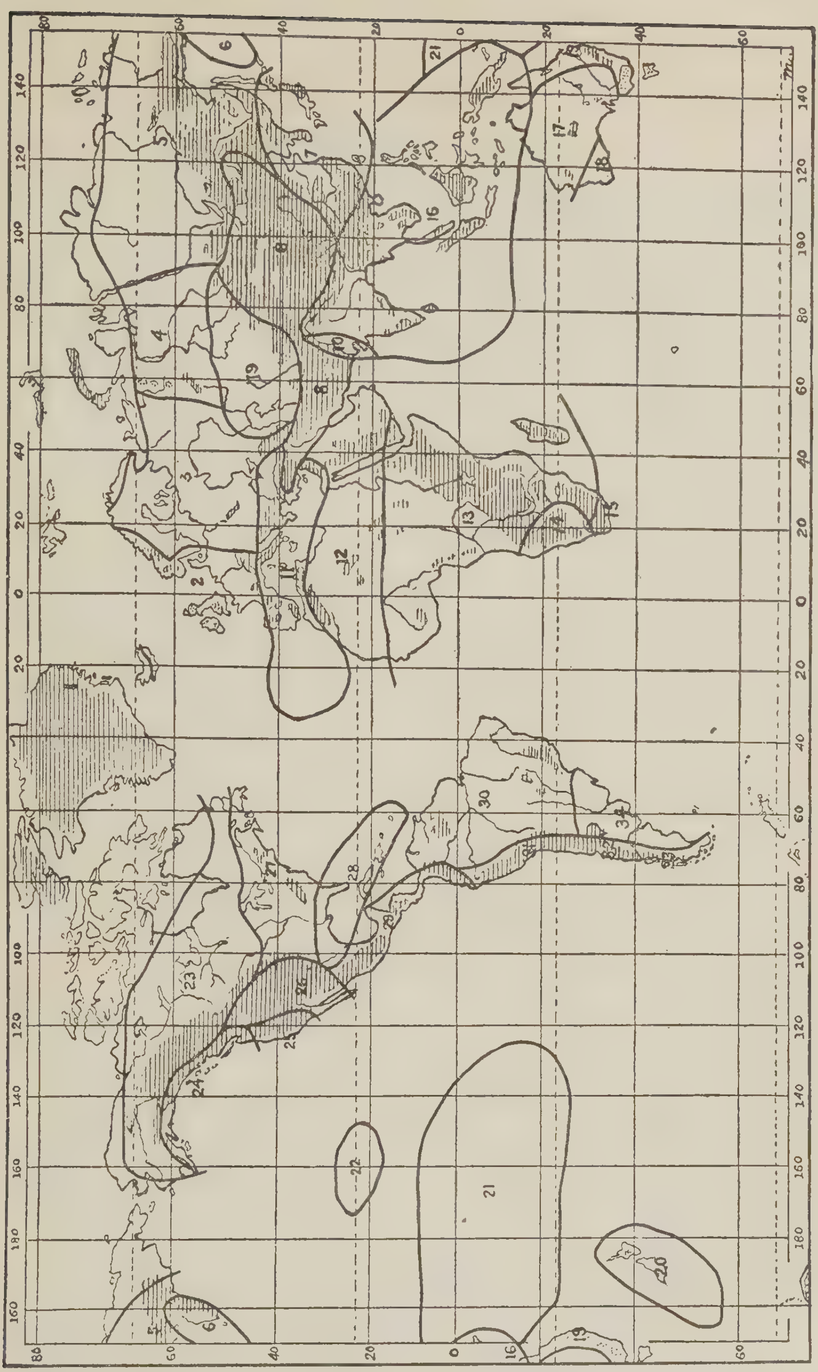

FIGURE 71. Supan's Climatic Provinces. The Plateau area above 3,000 feet is shown by shading. 1. Arctic. 2. West European. 3. East European. 4. West Siberian. 5. East Siberian. 6. Kamchatkan. 7. Sino-Japanese. 8. Asiatic Mountain and Plateau. 9. Aral. 10. Indus. 11. Mediterranean. 12. Saharan. 13. African Tropical. 14. Kalahari. 15. Cape. 16. Indo-Australian Monsoon. 17. Inner Australian. 18. Southwest Australian. 19. East Australian. 20. New Zealand. 21. Polynesian Tropical. 22. Hawaiian. 23. Hudson (North Canadian). 24. Northwest American Coastal. 25. California. 26. North American Mountain and Plateau. 27. Atlantic (East North American). 28. West Indian. 29. Tropical Cordilleran. 30. South American Tropical. 31. Peruvian. 32. North Chilean. 33. South Chilean. 34. Pampa.

CHANGES OF CLIMATE.

The geological record proves that vast changes of climate have occurred in the past, and there is every reason to suppose that such changes are still in progress. Even within historic times certain important climatic changes have been noted; especially the progressive desiccation or increased rainfall of various regions. Such changes are, however, exceedingly slow, and probably do not continue indefinitely in the same direction. There is evidence of oscillations of rainfall with periods of many centuries. Less extensive oscillations of climate and weather have also been tentatively established, and some of these appear to be connected with well-known fluctuations in solar activity. The spottedness of the sun's surface runs through a cycle averaging a little over 11 years. Astrophysical observations seem to show that at sunspot maximum more heat is given out by the sun than at sunspot minimum and accordingly it has been assumed that the earth's atmosphere as a whole is warmest when sunspots are most numerous, and *vice versa*. In the lower layers of the atmosphere, however, meteorological observations show that a reverse relation exists. In other words, air temperatures near the earth's surface average, for the world in general, somewhat lower at sunspot maximum than at sunspot minimum. This apparent contradiction has been attributed by some authorities to the fact that the effect of enhanced solar radiation is to increase the activity of the atmospheric circulation—thus increasing the flow of cold air, in the lower strata, from the poles toward the equator—and also to increase humidity, cloudiness and rainfall by promoting evaporation from the oceans. It is doubtful, however, whether this hypothesis can account for more than a brief initial cooling of the lower air due to increased insolation, and other explanations have been offered.[4] Both the 11-year sunspot period and other fluctuations in solar activity may be assumed *a priori* to be reflected, in some degree, in changes of terrestrial temperature, rainfall, storminess, etc., and many laborious studies have been devoted to these presumed relations, but definite conclusions cannot yet be formulated.

Investigators have also sought to discover weather periodicities from the study of meteorological and historical records, without reference to extra-terrestrial causes. The most celebrated investigations of this

[4] See especially the hypothesis of W. J. Humphreys, which takes account of probable changes in the spectral quality of solar radiation and their effects on the ozone content of the upper atmosphere. Jour. Franklin Institute, Sept., 1917, p. 408.

character are those of E. Brückner, who has accumulated much evidence in behalf of a period averaging about 35 years in the temperature, rainfall, and other meteorological elements, with important economic consequences (fluctuations in the prices of grain, etc.). A good summary of the literature on variations in weather and climate will be found in J. Hann's Handbuch der Klimatologie, vol. 1 (English translation of 2d edition, N. Y., 1903; but consult preferably the 3d German edition, Stuttgart, 1908).[5]

The universal popular belief in radical changes of climate within the period of a human lifetime—evinced, for example, in such expressions as "the old-fashioned winter"—has little if anything in common with the changes above discussed and is, for the most part, fallacious. This belief is chiefly due to the fact that extraordinary weather makes a more durable impression upon people's minds than ordinary weather. The illusion of the "old-fashioned winter" is also due, in many cases, to marked improvements in the heating of houses and the movements of population from the natural environment of the country to the artificial conditions of the town.

THE CLIMATE OF FRANCE.

Because of its present interest the following summary of the climate of France is reproduced from Handbook of Northern France, by William Morris Davis:

"The climate of France is much more temperate than the climate of an area of the same latitude in central or eastern North America. The prevailing winds come from the west and bring with them the tempering influences of the ocean; moreover, they come somewhat from the southwest in winter and thus diminish the cold, and somewhat from the northwest in summer and thus moderate the heat which would otherwise be felt. The mean temperature in January (from 6° C. = 43° F. in the south to 2° C. = 36° F. in the northeast) corresponds to that of North Carolina and northern Georgia or of Arkansas and Oklahoma in the same month. Winter weather is frequently cloudy and wet; hence the air is chilling though the temperature is not very low. The coldest winter winds are from the continental interior on the northeast. The mean temperature in July (from 24° C. = 75° F.

[5] See also the recent writings of W. J. Humphreys, Ellsworth Huntington and Henryk Arctowski, and an excellent review of the subject of solar-terrestrial relations in Helland-Hansen & Nansen's Temperatur-Schwankungen des Nordatlantischen Ozeans und in der Atmosphäre, Kristiania, 1917, p. 139 ffg.

in the southeast to 18° C. = 64° F. in the northwest) corresponds to the July mean of southern Pennsylvania and Ohio or of Wisconsin and North Dakota. The extremes of both seasons are less in France than in the central United States.

"The annual rainfall varies from 500 to 1000 millimeters (20 to 40 inches), corresponding in general terms to that of eastern Nebraska and Iowa. Snowfall is rarely heavy, even in the north; and as the winds that follow snowstorms usually come from the ocean at a temperature above freezing, snow seldom lies long on the ground. Weather changes, including the large cloudy areas of low barometric pressure with shifting winds and rain or snow, as well as the smaller thunderstorms of summer, advance in a general way from southwest to northeast, as in the eastern United States; but the tracks of low-pressure centers, which often traverse the United States, usually pass to the north of France in spite of its relatively high latitude; hence France more often receives the southerly than the northerly winds that spiral around such centers."

APPENDIX I.

LIST OF WORKS ON METEOROLOGY.

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Gockel, A. Die Luftelektrizität. Leipzig. 1908.
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Great Britain, Meteorological Office. The computer's handbook. London. 1916- [In course of publication, in parts.]
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Note: There are few meteorological libraries in America, and meteorology is not, as a rule, well represented in general libraries. By far the largest meteorological library in this country is that of the Weather Bureau in Washington. Information concerning books and papers on meteorology and its several branches can be obtained by addressing: Chief, U. S. Weather Bureau, Washington, D. C. Recent meteorological publications are listed regularly in the Monthly Weather Review.

APPENDIX II.

INTERNATIONAL METEOROLOGICAL SYMBOLS.

(For further descriptions of these symbols and an account of meteorological symbols in general see Monthly Weather Review, Wash., May, 1916, p. 265-274.)

Symbol.	Meaning.	Symbol.	Meaning.
	Rain.		Driving snow.
	Snow.	←	Ice-crystals.[3]
	Thunderstorm.		Snow on ground.
	Thunder.		Gale.
	Lightning.	⊙	Sunshine.
▲	Hail.	⊕	Solar halo.
△	Soft hail.	⦶	Solar corona.
≡	Fog.		Lunar halo.
	Ground fog.		Lunar corona.
	Wet fog.	⌒	Rainbow.
	Hoarfrost.		Aurora.
	Dew.		Zodiacal light.
V	Rime.[1]	∞	Haze.
	Glaze.[2]		

Exponents.—An exponent added to a symbol indicates the degree of intensity, ranging from ^{0} weak (light, etc.) to ^{2} strong (heavy, etc.). Thus, ^{0}, light rain; ^{2}, heavy rain. German and French observers use the exponent ^{1} to denote medium intensity. In English-speaking countries the omission of the exponent indicates medium intensity.

Time of occurrence.—When hours of occurrence are added to symbols, the abbreviation *a* is used for a.m., and *p* for p.m. Thus, 10a—4p denotes "rain from 10 a.m. to 4 p.m." The abbreviation *n* means "during night." Stations taking tri-daily observations may use *a* to mean the first and second observations; *p*, between the second and the third; and *n*, between the third and the first.

[1] A rough frost deposit from fog.

[2] A smooth coating of ice from cold rain; in Great Britain called "glazed frost" and in America, popularly, "sleet."

[3] Ice needles floating in the air.

Date Due